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THE COMPOSITION OF METEORITIC MATTER

III. PHASE EQUILIBRIA, GENETIC RELATIONSHIPS AND PLANET STRUCTURE

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ABSTRACT

An attempt has been made to study the composition of meteorites on a more quantitative basis than has been attempted heretofore. In particular, investigations have been made of experimental and theoretical approaches that might lead to more rigid comparisons between terrestrial and meteoritic matter. Stress has been placed particularly upon the following studies: (*a*) the distribution of elements between meteoritic phases; (*b*) average composition as a function of metal-phase content; (*c*) correlation between element distribution, thermochemical data, and general thermodynamic considerations.

It is demonstrated that if one assumes that the observed distributions of elements represent equilibrium distributions, then equilibrium must have been established at temperatures of the order of 3000° C. and pressures of the order of 10^5–10^6 atm. Similarly, it is demonstrated that the conditions at which equilibrium was achieved varied from meteorite to meteorite in such a way that the greater the metal-phase content, the greater the temperature and/or the pressure. The data indicate strongly that meteorites had their origin in a planet similar to the earth in general physicochemical characteristics.

INTRODUCTION

In 1850 A. Boisse suggested that the earth's composition is probably equivalent to the mean composition of meteoritic matter. Since that time much effort has been expended by astronomers, geologists, geophysicists, and, more recently, geochemists in attempts either to develop or to disprove Boisse's original speculation. On the whole, information accumulated during the last fifty years, notably by such men as Farrington, Merrill, Prior, Tammann, Goldschmidt, and Jeffreys, has served to substantiate in a qualitative way the thesis that meteorites belong to a single family possessing a common genesis, in all likelihood a planet similar in physicochemical characteristics to the earth. In particular, the investigations to date have resulted in the following conceptual advances: (1) belief that the dense core of the earth, as defined by seismic data, is probably similar to iron-nickel meteorites in composition; (2) belief that the silicate phase of stony meteorites is similar to the mantle surrounding the core of the earth; (3) realization that chemical equilibrium, or at least something approaching equilibrium, at one time existed between the various meteoritic phases; and (4) realization, in a qualitative way, of the physicochemical principles underlying the differentiation of elements during magma crystallization and resulting in the formation of igneous rocks from a mother-substance not too dissimilar in composition to the silicate phase of stony meteorites.

It must be emphasized that these

general comparisons have arisen almost entirely in a qualitative way, gross averages of meteoritic compositions having been used, in discussions both of the differentiation of elements during magma crystallization and of phase equilibria. More quantitative comparisons are highly desirable if the nature of the relationships, if any, between the earth and meteorites is to be ascertained and, looking still farther ahead, if the physicochemical processes that took place during the process of planet formation are to be understood, even partially.

This paper represents an attempt to study existing meteoritic data from a somewhat more quantitative point of view and to investigate experimental and theoretical approaches that might lead to more rigid comparisons between terrestrial and meteoritic matter.

THE AVERAGE COMPOSITION OF METEORITIC MATTER

The average composition of meteoritic matter tells us little about meteorites as a family. However, during the course of the present discussion, average compositions will be referred to from time to time; so, for convenience, the average compositions of the silicate phase of stony meteorites and iron meteorites and the gross composition of stony meteorites are given in table 1. The data given in columns 1, 2, and 3 are taken from recent recomputations by the authors (1947*a*, pp. 405–411; 1947*b*, pp. 508–510), and column 4 has been derived by combining the data of columns 1 and 2 with new computations of the average metal and sulphur contents of stony meteorites. The latter two averages are given in table 2, together with their standard deviations and precisions. Sulphur has been calculated as displacing oxygen when unaccounted for as FeS. The combined iron, shown in column 4 of table 1, represents the iron combined as oxide, sulphide, and phosphide.

The errors associated with the various components listed in column 4 of table 1 are greater, relative to one another, than in the other columns, owing largely to the substantial size of the standard deviation associated with the metal phase. The reasons for this unusually large standard deviation will become apparent when the metal-phase frequency curve is discussed later in the paper.

PHASE EQUILIBRIA IN METEORITES

Perhaps the most convincing evidence that fragments of meteoritic matter, now widely scattered, at one time existed in close contact with one another is the evidence derived from the distribution of elements between the various meteoritic phases. There are three main phases: the silicate phase, composed primarily of SiO_2, MgO, and FeO; the metal phase, composed primarily of iron, nickel, and cobalt; and troilite, composed primarily of FeS. In 1910 W. A. Wahl (p. 67) pointed out that those major meteoritic constituents whose oxides possess low heats of formation exist primarily in the metal phase and that those constituents whose oxides possess high heats of formation exist primarily in the silicate phase. The comprehensive studies by I. and W. Noddack (1930, p. 757; 1934, p. 173), V. M. Goldschmidt (1935, p. 183), and others on the distribution of minor constituents between the silicate and metal phases have well substantiated the fact that the phase (metal or silicate) in which a given element tends to concentrate depends markedly upon the affinity of the element for oxygen.

Prior (1916, p. 26), surveying meteoritic compositions more closely, observed that, in general, the smaller the

concentration of metal phase in a stony meteorite, the greater the concentration of nickel in the metal phase and the greater the concentration of combined iron in the silicate phase. Prior (1920, p. 54) advanced the explanation that "meteorites have separated from a single magma which has passed through successive stages of progressive oxidation."

and that, in the absence of sulphur, the equilibrium constant is independent of the composition of the silicate phase. He found (1931, p. 46) that sulphur slightly displaces the equilibrium. The constants as measured by Zur Strassen at two temperatures in the absence of sulphur are given in table 3.

The observed "average" equilibrium constant for iron and nickel in stony

TABLE 1

THE AVERAGE COMPOSITION OF STONY AND IRON METEORITES
(Per Cent by Weight)

Element	1 Silicate Phase of Stones	2 Metal Phase of Stones	3 Iron Meteorites	4 Total Stony Meteorites
Oxygen	43.12±1.00			36.15±0.89
Silicon	21.61±0.19			18.12±0.22
Comb. iron	13.23±0.42			14.21±0.39
Comb. nickel	0.39±0.07			0.33±0.06
Comb. cobalt	0.02±0.01			0.02±0.01
Magnesium	16.62±0.32			13.93±0.29
Sulphur				1.79±0.08
Calcium	2.07±0.21			1.74±0.18
Aluminum	1.83±0.15			1.53±0.13
Sodium	0.82±0.06			0.69±0.05
Chromium	0.36±0.06			0.30±0.05
Manganese	0.31±0.07			0.26±0.06
Potassium	0.21±0.02			0.18±0.02
Phosphorus	0.17±0.01			0.14±0.01
Titanium	0.10±0.03			0.08±0.03
Hydrogen	0.07±0.02			0.06±0.02
Metallic iron		88.58±0.55	90.78±0.26	9.97±0.69
Metallic nickel		10.68±0.51	8.59±0.24	1.20±0.10
Metallic cobalt		0.70±0.06	0.63±0.02	0.08±0.01

The $Ni–FeSiO_3–NiSiO_3–Fe$ equilibrium has been studied by Zur Strassen (1930, p. 209), who measured the equilibrium constant for the reaction

$$Ni + FeSiO_3 \rightleftarrows Fe + NiSiO_3 ,$$

where the constant C is given by the expression

$$C = \frac{(Fe)_m\,(Ni)_{si}}{(Ni)_m\,(Fe)_{si}}.$$

Zur Strassen found that the reaction obeys the mass-action law over a wide range of iron and nickel compositions

TABLE 2

CONCENTRATION OF METAL PHASE AND SULPHUR IN STONY METEORITES

Substance	Average Concentration (Per Cent)	Standard Deviation (Per Cent)	Number of Cases	Precision (Per Cent)
Metal phase	11.25	8.94	130	0.78
Sulphur	1.79	0.93	130	0.08

meteorites as calculated from table 1 is

$$C = 0.24 ,$$

a value nearly forty times greater than

Zur Strassen's measured value at 1750° K. A study of the effects of sulphur serves to demonstrate that the average sulphur content of stony meteorites (1.79 per cent by weight) is too small to produce any appreciable effect upon the equilibrium under the conditions used by Zur Strassen. One is thus forced to the conclusion that, if equilibrium existed between the metal and silicate phases of stony meteorites, it must have been established under conditions of temperature or pressure markedly different from 1750° K. and 1 atm.

TABLE 3

MEASURED EQUILIBRIUM CONSTANTS FOR THE REACTION $Ni + FeSiO_3 \rightleftarrows NiSiO_3 + Fe$
(Data of Zur Strassen [1930])

Temperature ° K.	C	Remarks
1750	6.53×10^{-3}	Average
1840	7.25×10^{-3}	Average
1840	7.44×10^{-3}	Highest value

That the temperature effect is in a direction indicating that equilibrium took place at a temperature higher than 1750° K. can be seen from the fact that the reaction is endothermic and by applying the relationship[1]

$$\frac{d \ln K}{dT} = \frac{\Delta H}{RT^2} \quad (\Delta H = \text{The enthalpy change for the reaction}),$$

thus obtaining a positive value for the change of lnK with increasing temperature. However, preliminary calculations, to be presented at a later date, indicate that the high observed value of the "average" equilibrium constant in meteorites cannot be easily accounted for on an increased-temperature basis alone, the temperature obtained being unreasonably high.

That increasing pressures would tend to push the equilibirum concentration toward the nickel silicate side can be seen from an inspection of the relative volumes of iron and nickel metals and oxides, indicating that a volume decrease probably results from the silicate reaction. The change in volume is illustrated in table 4.[2]

TABLE 4

VOLUME CHANGE IN THE REACTION
$Ni + FeO \rightarrow NiO + Fe$

Substance	Molar Volume (cc.)
Ni	6.59
FeO	12.60
NiO	10.03
Fe	7.07
Change	− 2.09

It is interesting to investigate the order of magnitude of the pressure that would be required in order to produce an appreciable shift in the equilibrium. The volume decrease per mole of reactants is of the order of 2 cc. We then have, for the free-energy charge in increasing the pressure at constant temperature:[3]

[1] The thermodynamic nomenclature used in this discussion is that of Lewis and Randall (1923).

[2] A similar calculation based upon the densities of the silicates $2FeO \cdot SiO_2$ (Roth and Troitzsch, 1932–33, p. 79) and $2NiO \cdot SiO_2$ (Taylor, 1930, p. 241) gives a volume decrease of 2.27 cc. per mole of reactant metals.

[3] The symbol K is used in this discussion to denote the true equilibrium constant. For the purposes of approximate calculation it is assumed that the mass-action constant, C, lies close to the true equilibrium constant.

$$\Delta F = RT ln \frac{K_{obs}}{K_{p=0}} \qquad K_{obs} = \text{Observed equilibrium constant for meteorites;}$$
$$K_{p=0} = \text{Equilibrium constant at zero pressure;}$$
$$\cong RT ln\ 40 .$$

For a temperature of 2000° K. this corresponds to an energy change of 7.4 K cal/mole. Remembering that $\Delta F = \int \Delta V dP$ and that $\Delta V = 2$ cc/mole, we have

$$P \cong 1.5 \times 10^5 \text{ atm.}$$

Such a pressure is comparable to the internal pressures existing within Mars.

A more informative picture of the relationships between iron, nickel, and their silicates within meteorites results from a study of the iron and nickel contents of stony meteorites as a function of their metal-phase contents. The authors (1947*b*, pp. 508–510) have recently confirmed Prior's relationship concerning the nickel content of the metal phase by the process of arranging a series of ninety-five selected stony meteorite analyses in order of increasing metal-phase content and then averaging the nickel contents of the metal phases over specified intervals of metal-phase content. The results, reprinted from the authors' previous paper, are shown in figure 1. If now, instead of nickel concentrations, the average equilibrium constant within each range is plotted as a function of metal-phase concentration, the result shown in figure 2 is obtained.[4] From figure 2 it can be seen that as the metal-phase content increases, the equilibrium constant increases.

On the basis of the curve shown in figure 2, one must conclude that, if the iron and nickel concentrations in stony meteorites represent equilibrium concentrations, then *the conditions under which equilibrium was achieved varied from meteorite to meteorite in such a way that the greater the metal-phase content, the higher the temperature or the pressure.*

Additional information supporting the possibility that equilibrium at one time existed between the silicate and the metal phases can be obtained from a close inspection of the distribution of a series of elements between the two phases and by a comparison of the distribution coefficients with the heats of formation of the oxides (in the absence of adequate thermal data on the silicates). One would expect that if the heat of the reaction is the prime contributor to the free energy, then an approximation to a straight line would result when

$$\log \frac{(\text{Fe})_m (M)_{si}}{(M_m)\ (\text{Fe})_{si}}$$

is plotted against the differences between the heats of formation of the metal oxides (per oxygen atom) and ferrous oxide.

Table 5 gives the data for the distribution of several elements between the silicate and metal phases, together with the heats of formation of their oxides. The results are plotted in figure 3. It can be seen that a straight line yielding a temperature of approximately 3000° C. is not an unreasonable approximation to the data, although the crudeness of such an estimation cannot be overemphasized. Here, as in the case of nickel alone, if pressure effects are, in reality, of considerable importance in establishing equilibrium, the shape of the curve shown in

[4] Unfortunately, there are very few analyses of the silicate phase of stony meteorites in which the NiO results can be trusted. However, a sufficient number of analyses has been accumulated to permit each point in fig. 2 to be represented by the average NiO content of three or more stony meteorites.

figure 3 and consequently the calculated temperature would be considerably altered.

In view of the strong indication that the equilibrium distribution of elements between the metal and silicate phases depends in large measure upon the relative heats of formation of the oxides (or silicates) involved, one might expect that the distribution of elements between the sulphide and metal phases would depend in like manner upon the relative heats of formation of the sulphides involved. Actually, close inspection of the meteoritic sulphide-metal distribution coefficients and thermal data reveals little or no correlation between the two quantities. However, if we compare the relative heats of formation of oxides and sulphides (table 6) and once again assume the existence of a pressure effect in establishing equilibrium, an explanation presents itself.

It can be seen from table 6 that the heats of formation of sulphides are considerably lower than the heats of formation of the corresponding oxides. As a result, the energy change in a reaction of the type

$$M + FeS \rightleftharpoons MS + Fe$$

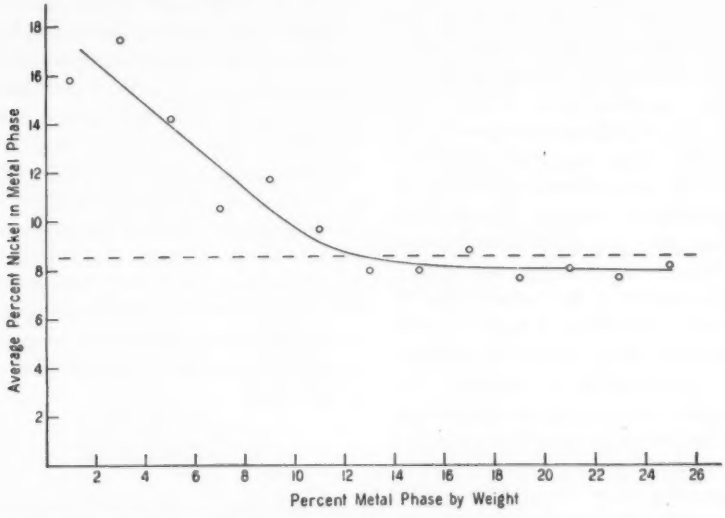

FIG. 1.—The variation of the nickel content in the metal phase of stony meteorites with the metal-phase content in the meteorite. The dotted line represents the average nickel content of iron meteorites. Each point represents the average of between two and nineteen stony meteorites.

will generally be considerably smaller than the energy change in a reaction of the type

$$M + FeSiO_3 \rightleftharpoons MSiO_3 + Fe\,.$$

Recalling the indications that a pressure-volume effect is required to explain the large equilibrium constant in the case of the iron-nickel-silicate equilibrium and remembering that in such a case $P\Delta V$ would be of the same order of magnitude as the chemical energy change, we see

that *in the case of the sulphides, where the heats of formation are small,* $P\Delta V$ *would be the major factor in establishing equilibrium.*

Reference once again to the case of nickel can elucidate the point. Zur Strassen (1931) has measured equilibrium constants for the reaction.

$$Ni + FeS \rightleftharpoons NiS + Fe ,$$

and at about 1200° K. he found that the constant C possesses a value very close to unity. If we now look at the observed distribution of nickel and iron between troilite and metal, we find that for meteorites

$$C \cong 0.02 ,$$

a value so low as to be practically unexplainable on a temperature-difference

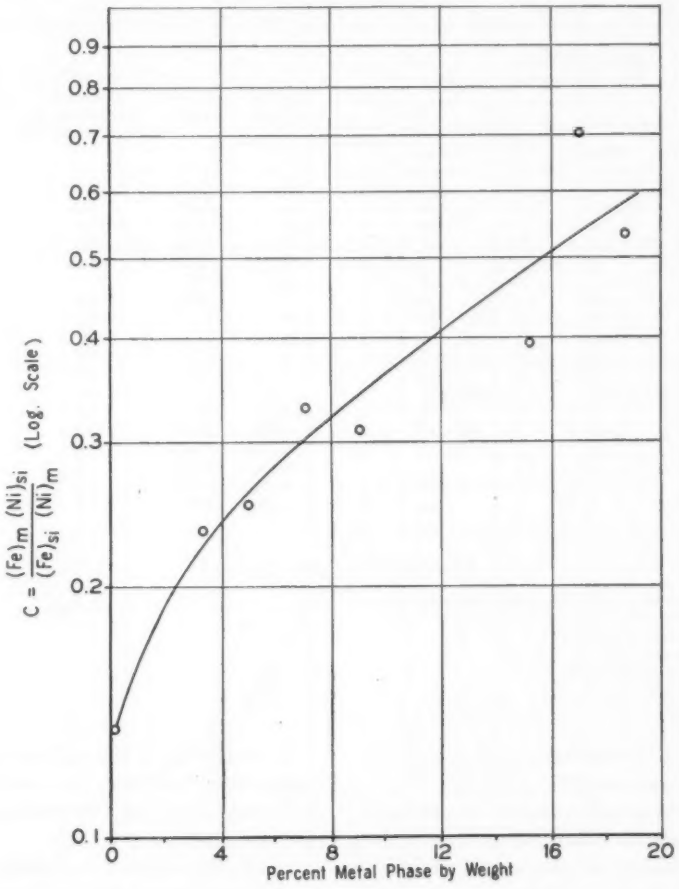

FIG. 2.—The iron-nickel equilibrium constant, $C = \frac{(Fe)_m(Ni)_{si}}{(Ni)_m(Fe)_{si}}$, in stony meteorites as a function of the metal-phase content.

basis. However, if the molar volumes of the components of the reaction are examined (table 7), it will be seen that a volume increase is associated with the reaction as it proceeds to the right. A pressure effect would, as a consequence, shift the equilibrium to the left.

definite proof must await refined thermochemical calculations, more adequate thermal data, more precise measurements of distribution coefficients, and a broader theoretical foundation of the effects of high pressures on partial molal volumes.

TABLE 5

HEATS OF FORMATION AND METEORITIC DISTRIBUTION COEFFICIENTS OF VARIOUS METAL OXIDES

ELEMENT	SILICATE PHASE OF STONY METEORITES		IRON METEORITES		$\frac{(Fe)_m(M)_{si}}{(Fe)_{si}(M)_m}$	COMPOUND	ΔH OF FORMATION PER OXYGEN ATOM (K CAL/MOLE)	ΔH OF FORMATION PER OXYGEN ATOM MINUS $(\Delta H)_{FeO}$
	Parts per Million	Ref.	Parts per Million	Ref.				
Ca	20,700	1	500	3	280	CaO	152	86.3
Mg	166,000	1	320	3	3,500	MgO	144	78.3
Al	18,300	1	40	3	3,100	Al_2O_3	133	67.3
Ti	1,000	1	100	4	68	TiO_2	109	43.3
V	50	5	6.2	3	55	V_2O_5	105	39.3
Si	216,000	1	8,000	3	180	SiO_2	101	35.3
Mn	3,100	1	300	3	70	MnO	90.8	25.1
Cr	3,900	5	300	5	88	Cr_2O_3	89.3	23.6
Zr	100	3	8	3	85	ZrO_2	89.3	23.6
Zn	76	6	115	3	4.5	ZnO	85	19.3
Sn	4	3	102	3	0.27	SnO	69.8	4.1
Fe	132,000	1	897,930	2	1.0	FeO	65.7	0
Mo	2.5	3	16.6	3	1.0	MoO_2	65.7	0
Ni	3,900	1	84,947	2	0.31	NiO	57.9	− 7.8
Co	200	1	6,230	2	0.22	CoO	57.5	− 8.2
Pb	2	5	53	3	0.26	PbO	52.5	−13.2
Cu	1.6	3	305	3	0.036	CuO	35	−30.7
Pt	0.08	3	19.4	6	0.029	PtO	17	−48.7

REFERENCES FOR TABLE 5

1. Brown and Patterson, 1947*a*, pp. 405–411.
2. Brown and Patterson, 1947*b*, pp. 508–510.
3. Noddack, 1930, p. 757.
4. Ishibashi, 1931, p. 372.
5. Goldschmidt, 1938.
6. Noddack, 1934, p. 173.

It must be emphasized that any conclusions concerning the existence of pressure effects as main factors in determining meteoritic distribution coefficients are predicated on the assumption that the coefficients as observed are somewhat close to equilibrium values. Although existing data, such as that shown in figure 2, support such an assumption,

Nevertheless, if the validity of the assumption is admitted, the conclusions to be drawn on the basis of existing data appear straightforward:

1. Equilibrium was established under conditions of relatively high temperature and pressure (approximately 3000° C. and 10^5–10^6 atm.).

2. The temperature and pressure ef-

fects were such that the greater the metal content of a fragment of meteoritic matter, the higher the temperature and/or the pressure at which equilibrium was attained.

In the light of these conclusions, the temptation becomes great to go further and to associate meteoritic matter with a planet in which the metal content, the temperature, and the pressure increased toward the center. This, of course, is much the same conclusion as that drawn so many years ago by Boisse; and, as will be shown below, the more closely one examines the data, the more clearly does such a conclusion appear to be justified.

GENETIC RELATIONSHIPS

Of all major meteoritic constituents, the one which varies over the widest

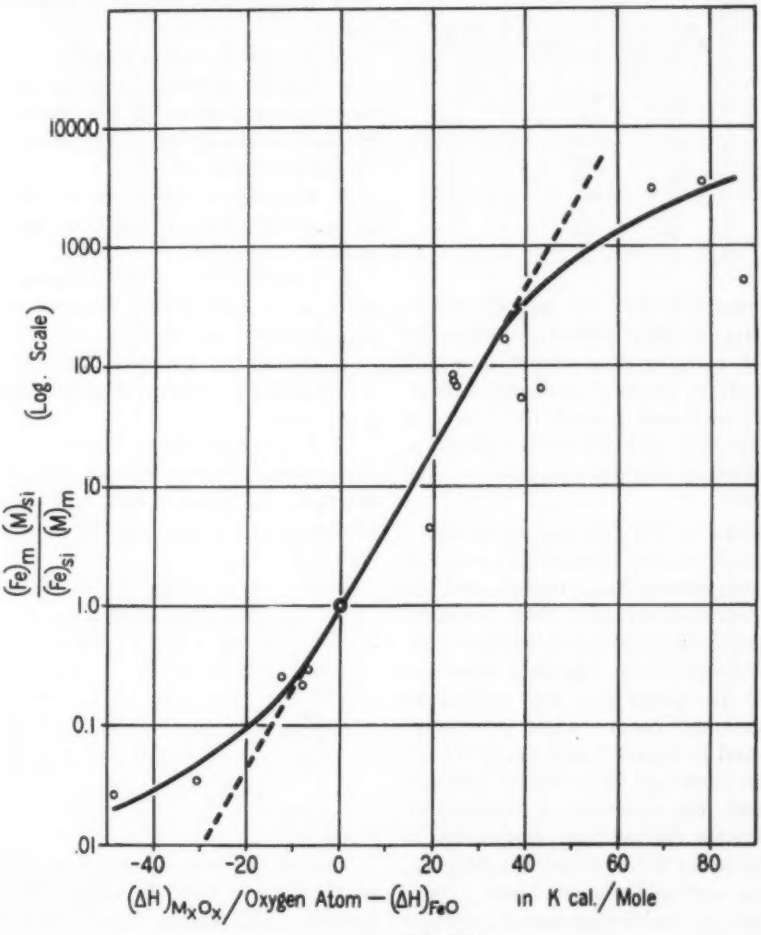

FIG. 3.—Observed meteoritic equilibrium constants, $\frac{(Fe)_m(M)_{si}}{(Fe)_{si}(M)_m}$, for various elements as a function of the heats of formation of their oxides.

range is metallic iron. Many meteorites are known in which the metal-phase content is close to zero; many are known in which the metal-phase content is 100 per cent; and specimens have been found representing every gradation between these two extremes. If, as our phase-equilibria studies indicate, meteorites actually came from a planet in which temperature, pressure, and metal-phase content increased toward the center, a study of gross meteoritic composition as a function of metal-phase content is of interest.

TABLE 6

HEATS OF FORMATION OF A FEW METAL OXIDES AND SULPHIDES

Compound	Heat of Formation (K Cal/Mole)	Compound	Heat of Formation (K Cal/Mole)
FeO	65.7	FeS	23.1
NiO	57.9	NiS	17.4
CoO	57.5	CoS	19.7
CuO	34.9	CuS	11.6
ZnO	85.0	ZnS	45.9
SnO	69.8	SnS	23.9
CdO	65.2	CdS	33.9
PbO	52.5	PbS	22.2

A series of 127 selected stony-meteorite analyses was arranged in order of increasing metal-phase content, and the gross composition was then averaged over specific intervals of metal-phase content, thus obtaining the data shown in table 8. In figures 4, 5, and 6 the data are presented on a weight percentage basis, and in figures 7 and 8 some of the data are presented on an atomic percentage basis. For convenience, corresponding data for the average composition of igneous rocks (Clarke and Washington, 1922, p. 108) and plateau basalt (Daly, 1933, pp. 17, 201) are presented, together with the meteoritic data.

From the figures several interesting features become apparent. As the metal-phase content increases:

1. Combined iron[5] increases, passing through a maximum at a metal-phase content of 3–4 per cent, following which the combined iron decreases steadily.

2. Total iron (and total iron, nickel, and cobalt) increases rapidly until a metal-phase content of 3–4 per cent is reached, following which the increase with metal-phase content is more gradual.

3. Oxygen decreases steadily.

4. Silicon decreases rapidly until a metal-phase content of 3–4 per cent is reached, following which the silicon concentration levels off.

5. Magnesium increases, leveling off at a metal-phase content of approximately 5 per cent.

6. Calcium decreases, becoming constant at a metal-phase content of approximately 5 per cent.

7. Aluminum decreases, leveling off at a metal-phase content of approximately 5 per cent.

8. Sulphur increases rapidly until a metal-phase content of 3–4 per cent is reached, following which the general tendency is for the sulphur to remain constant.

TABLE 7

VOLUME CHANGE IN THE REACTION $Ni + FeS \rightleftarrows NiS + Fe$

Substance	Molar Volume (cc.)
Ni	6.59
FeS	18.16
NiS	19.73
Fe	7.07
Change	+ 2.05

One of the more interesting features of the data is the relationship between igneous rock, plateau basalt, and stony

[5] "Combined iron" is iron combined as oxides and sulphides.

TABLE 8

AVERAGE COMPOSITION OF STONY METEORITES AS A FUNCTION OF METAL-PHASE CONTENT
("Wt. %" Denotes Per Cent by Weight. "A%" Denotes Atomic Per Cent)

Metal-Phase Interval Per Cent by Weight		Combined Iron	Metallic Iron	Total Iron	Combined Nickel	Total Nickel	Total Cobalt	Total Iron, Nickel, Cobalt	Oxygen	Silicon	Magnesium	Aluminum	Calcium	Sulphur	Sodium	Potassium	Titanium	Phosphorus	Manganese	Chromium	Metal Phase	Number of Analyses
Igneous rock	Wt. %	5.10	.....	5.10	0.01	0.01	.....	5.11	40.49	27.62	2.11	8.12	3.63	0.05	2.85	2.60	0.63	0.13	0.09	0.04	.....	.....
	A%	1.94	.....	1.94	.....	.....	.....	1.94	61.83	20.95	1.85	6.41	1.93	.....	2.64	1.42	0.28	0.09	0.03	.....	.....	
Plateau basalt	Wt. %	8.09	.....	8.09	.....	.....	.....	8.09	45.41	23.64	2.95	9.21	5.78	.....	2.40	1.46	0.41	0.09	0.36	.....	.....	.....
	A%	3.10	.....	3.10	.....	.....	.....	3.10	60.78	18.04	2.60	7.31	3.09	.....	2.24	0.80	0.18	0.06	0.14	.....	.....	
0– 1	Wt. %	14.04	0.19	14.23	0.37	0.41	0.04	14.69	41.58	21.92	10.93	4.31	6.08	0.61	0.51	0.12	0.40	0.07	0.27	0.28	0.23	14
	A%	5.54	0.08	5.61	.....	0.15	0.02	5.78	57.26	17.21	9.90	3.52	3.34	0.42	0.49	0.07	0.18	0.05	0.11	0.12	.....	
1– 2	Wt. %	14.79	1.06	15.85	.....	0.21	0.02	16.08	41.39	21.87	12.22	2.98	3.29	1.01	0.47	0.19	0.55	0.05	0.23	0.28	1.29	3
	A%	5.97	0.43	6.41	.....	0.08	0.01	6.50	58.41	17.00	11.35	2.50	1.85	0.71	0.46	0.11	0.26	0.04	0.09	0.12	.....	
2– 3	Wt. %	17.63	1.87	19.50	0.05	1.05	0.21	20.66	38.70	19.10	16.04	1.13	1.26	1.75	0.63	0.08	.....	0.10	0.22	0.25	2.30	5
	A%	7.34	0.78	8.12	.....	0.42	0.08	8.62	56.26	15.83	15.34	0.97	0.73	1.27	0.64	0.05	.....	0.07	0.09	0.11	.....	
3– 4	Wt. %	20.67	2.78	23.45	0.84	1.44	0.10	24.99	37.78	18.00	11.86	1.34	2.16	2.14	0.80	0.18	.....	0.29	0.42	0.42	3.43	4
	A%	8.87	1.19	10.07	.....	0.59	0.04	10.69	56.60	15.38	11.69	1.19	1.29	1.60	0.83	0.11	.....	0.23	0.18	0.19	.....	
4– 6	Wt. %	16.13	4.28	20.41	0.55	1.27	0.13	21.81	38.76	17.70	15.24	2.21	1.39	1.73	0.68	0.20	0.10	0.10	0.16	0.31	5.05	14
	A%	6.73	1.79	8.52	.....	0.50	0.05	9.08	56.49	14.71	14.61	1.91	0.81	1.26	0.69	0.12	0.05	0.07	0.07	0.14	.....	
6– 8	Wt. %	16.43	6.33	22.76	0.51	1.26	0.08	24.10	38.03	18.37	14.86	1.29	1.16	1.88	0.69	0.14	0.11	0.17	0.17	0.27	7.13	11
	A%	6.88	2.65	9.53	.....	0.50	0.03	10.07	55.99	15.31	14.29	1.12	0.68	1.37	0.70	0.08	0.05	0.13	0.07	0.12	.....	

TABLE 8—*Continued*

Metal-Phase Interval Per Cent by Weight		Combined Iron	Metallic Iron	Total Iron	Combined Nickel	Total Nickel	Total Cobalt	Total Iron, Nickel, Cobalt	Oxygen	Silicon	Magnesium	Aluminum	Calcium	Sulphur	Sodium	Potassium	Titanium	Phosphorus	Manganese	Chromium	Metal Phase	Number of Analyses
8–10	Wt.%	15.06	7.96	23.02	0.46	1.53	0.11	24.66	36.27	18.40	14.24	1.28	1.40	2.26	0.73	0.20	0.09	0.14	0.40	0.33	9.09	19
	A%	6.48	3.43	9.91	.....	0.63	0.05	10.59	54.52	15.77	14.08	1.14	0.84	1.70	0.76	0.12	0.05	0.11	0.18	0.15	.....	
10–12	Wt.%	14.16	9.97	24.13	0.15	1.22	0.08	25.43	36.44	18.36	14.38	1.32	1.23	1.83	0.56	0.33	0.04	0.16	0.29	0.26	11.10	8
	A%	6.09	4.29	10.38	.....	0.50	0.03	10.91	54.71	15.72	14.20	1.17	0.74	1.37	0.58	0.20	0.02	0.12	0.13	0.12	.....	
12–14	Wt.%	13.29	11.90	25.19	0.24	1.28	0.09	26.56	35.68	17.92	14.32	1.62	1.28	1.57	0.97	0.13	0.02	0.07	0.20	0.35	13.03	6
	A%	5.76	5.16	10.92	.....	0.53	0.04	11.48	53.98	15.46	14.25	1.45	0.77	1.19	1.02	0.08	0.01	0.06	0.09	0.16	.....	
14–16	Wt.%	13.36	13.95	27.31	0.33	1.56	0.10	28.97	33.73	17.59	13.23	1.88	1.46	2.19	0.56	0.35	0.02	0.19	0.58	0.26	15.24	6
	A%	5.94	6.20	12.15	.....	0.66	0.04	12.85	52.36	15.57	13.51	1.73	0.90	1.70	0.61	0.22	0.01	0.15	0.26	0.12	.....	
16–18	Wt.%	11.77	15.49	27.26	0.56	2.05	0.13	29.44	33.47	16.85	13.74	1.51	1.39	2.02	1.04	0.23	0.06	0.12	0.76	0.32	17.06	10
	A%	5.25	6.91	12.16	.....	0.87	0.05	13.09	52.14	14.97	14.08	1.40	0.86	1.57	1.13	0.15	0.03	0.10	0.34	0.15	.....	
18–20	Wt.%	10.46	17.18	27.64	0.32	1.75	0.11	29.50	33.51	17.12	14.34	1.33	1.29	1.85	0.73	0.12	0.12	0.13	0.20	0.20	18.72	13
	A%	4.67	7.67	12.35	.....	0.74	0.05	13.14	52.25	15.22	14.71	1.23	0.80	1.44	0.79	0.08	0.06	0.10	0.09	0.09	.....	
20–22	Wt.%	9.45	18.93	28.38	0.08	1.74	0.09	30.21	32.80	17.10	14.07	1.32	1.46	2.03	0.59	0.23	0.10	0.11	0.19	0.22	20.67	7
	A%	4.26	8.54	12.80	....	0.75	0.04	13.58	51.63	15.35	14.57	1.23	0.92	1.59	0.65	0.15	0.05	0.09	0.09	0.11	.....	
22–24	Wt.%	9.14	21.16	30.30	0.16	1.94	0.12	32.36	31.52	16.08	13.97	1.51	1.31	2.14	0.69	0.12	0.05	0.08	0.26	0.18	23.06	4
	A%	4.20	9.73	13.93		0.85	0.05	14.83	50.57	14.71	14.74	1.44	0.84	1.71	0.77	0.08	0.03	0.07	0.12	0.09		
24–26	Wt.%	8.90	22.48	31.38	0.09	2.09	0.13	33.60	30.71	15.90	13.71	1.24	1.37	2.48	0.78	0.16	0.05	0.03	0.30	0.18	24.61	3
	A%	4.13	10.44	14.57	.....	0.92	0.06	15.55	49.78	14.69	14.62	1.19	0.89	2.01	0.88	0.11	0.03	0.03	0.14	0.09	.....	

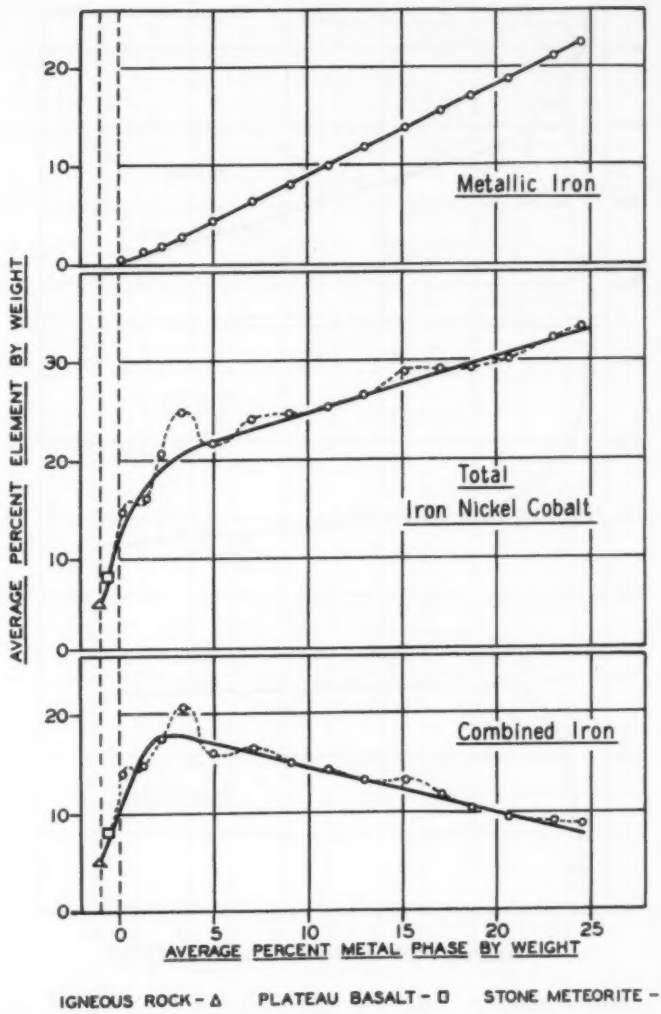

FIG. 4.—Variation of the weight percentages of metallic iron, total iron, nickel, and cobalt and combined iron in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

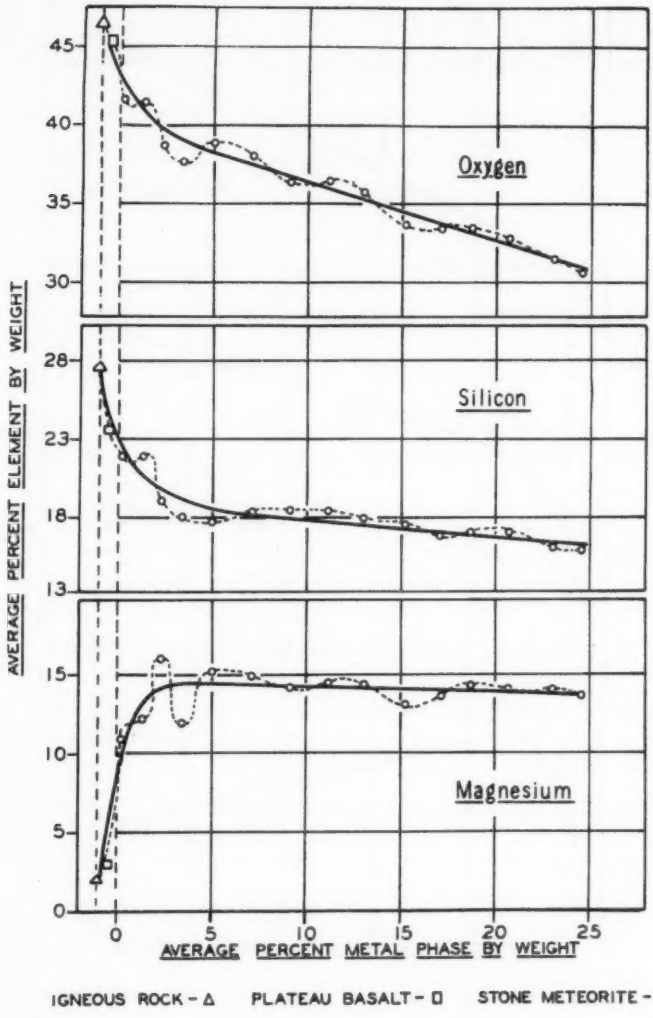

FIG. 5.—Variation of the weight percentages of oxygen, silicon, and magnesium in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

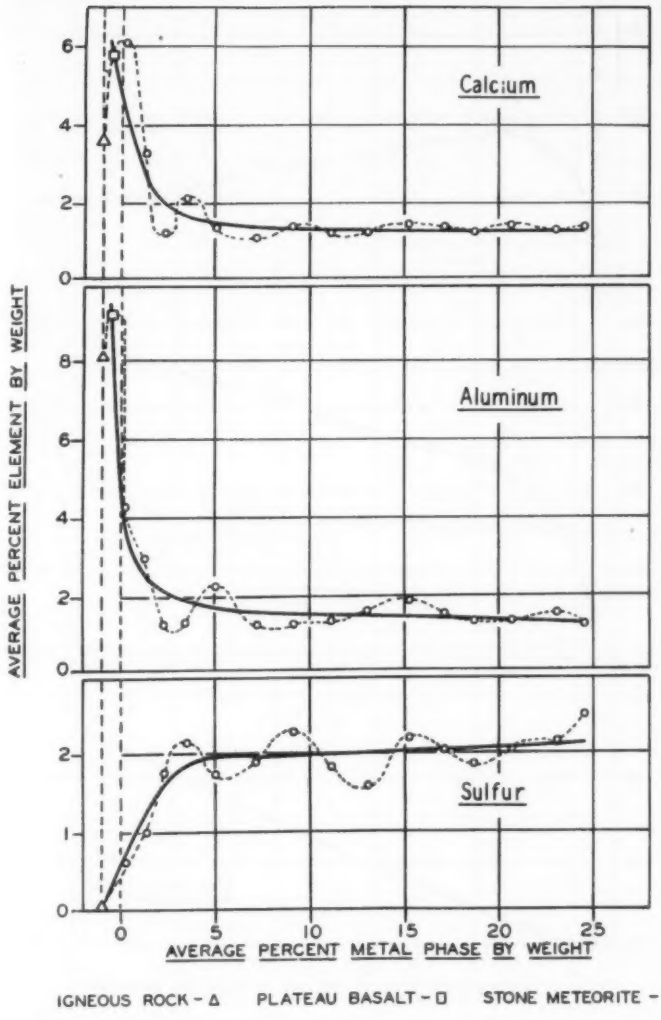

FIG. 6.—Variation of the weight percentages of calcium, aluminum, and sulphur in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparison.

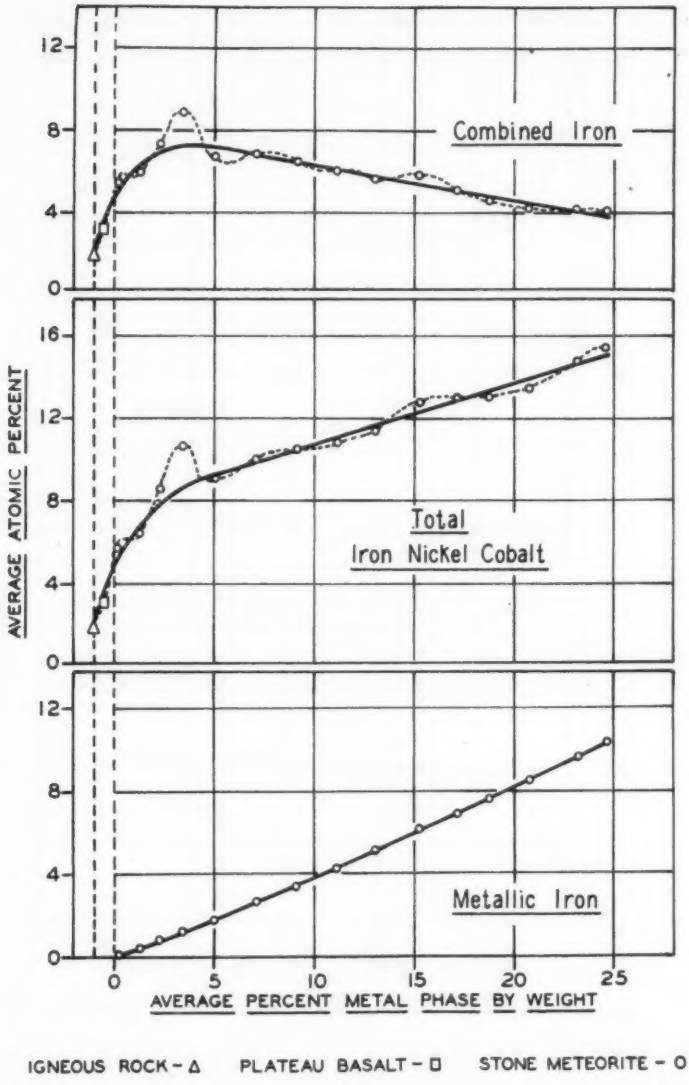
Combined Iron
Total
Iron Nickel Cobalt
Metallic Iron
AVERAGE ATOMIC PERCENT
AVERAGE PERCENT METAL PHASE BY WEIGHT
IGNEOUS ROCK - Δ PLATEAU BASALT - □ STONE METEORITE - O

FIG. 7.—Variation of the atomic percentages of combined iron, total iron, nickel, and cobalt and metal iron in stony meteorites as a function of metal-phase content. Earth's-crust values are given for comparis

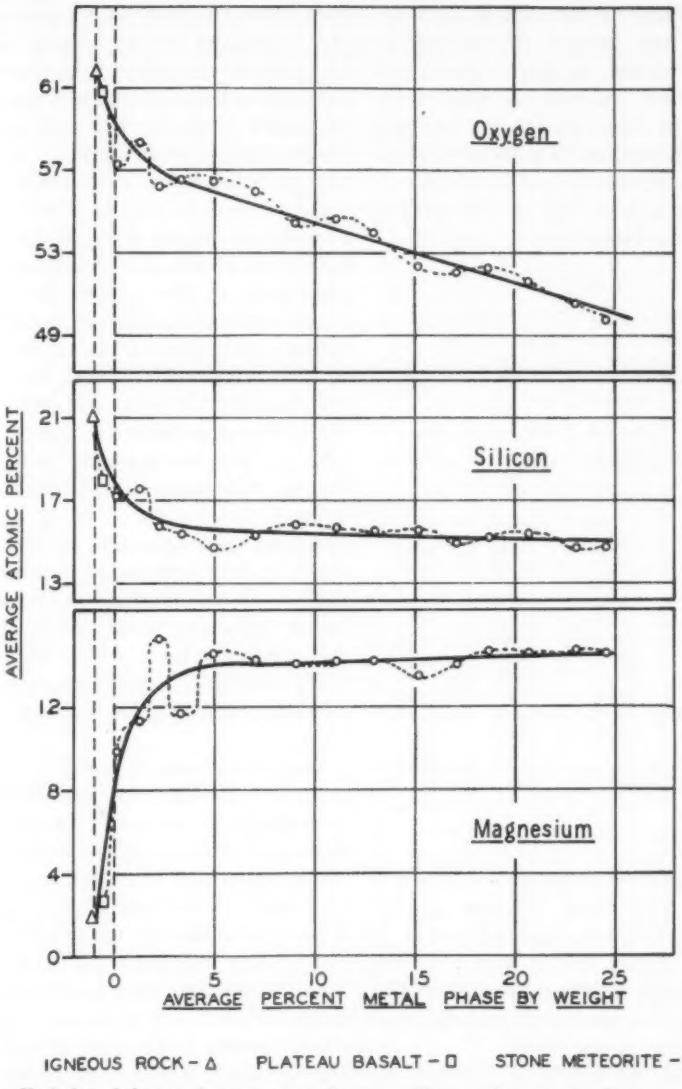

FIG. 8.—Variation of the atomic percentages of oxygen, silicon, and magnesium in stony meteorites as function of metal-phase content. Earth's-crust values are given for comparison.

meteorites which possess low metal-phase contents. *In the cases of combined iron, total iron, oxygen, silicon, magnesium, and calcium, the plateau basalt values are clearly intermediate between the corresponding values for igneous rock and meteorites.* Aluminum is a definite exception to this regularity; and in the case of sulphur the lack of data for plateau basalt prevents comparison.

TABLE 9

RELATIVE ABUNDANCE OF TRANSITION METALS IN STONY METEORITES

Metal-Phase Range (Weight Per Cent)	Av. Total Iron, Nickel, Cobalt (Atomic Per Cent)	Av. Oxygen (Atomic Per Cent)	Sum of Avs.
0– 1	5.78	57.26	63.04
1– 2	6.50	58.41	64.91
2– 3	8.62	56.20	64.88
3– 4	10.69	56.60	67.29
4– 6	9.08	56.49	65.57
6– 8	10.07	55.99	66.06
8–10	10.59	54.52	65.11
10–12	10.91	54.71	65.62
12–14	11.48	53.98	65.46
14–16	12.85	52.36	65.21
16–18	13.09	52.14	65.23
18–20	13.14	52.25	65.39
20–22	13.58	51.63	65.21
22–24	14.83	50.57	65.40
24–26	15.55	49.78	65.33

Indeed, it seems improbable that such a marked relationship between igneous rock, plateau basalt, and meteorites is fortuitous. Again one is forced to the conclusion that meteorites once formed a planet similar in general physicochemical characteristics to the earth.

Inspection of figures 4–8 serves to show that at metal-phase contents greater than 4–5 per cent, the main variables in meteorites are iron and oxygen, with the decrease in oxygen almost exactly matching the increase in iron, cobalt, and nickel, on an atomic percentage basis. This relationship between total iron, cobalt, and nickel and oxygen is of remarkable constancy, as can be seen from table 9, where the sums of the atomic percentages of these metals and oxygen are tabulated. It seems clear that at metal-phase contents greater than 4–5 per cent, the gradation from one meteorite to another consists primarily of the removal of atoms of oxygen (from atoms of combined iron, cobalt, and nickel) and the replacement of the oxygen atoms with equal numbers of atoms of metallic iron, cobalt, and nickel. Of course, as a result of this relationship between oxygen and the transition elements, the ratio of combined iron to metallic iron increases rapidly as the metal-phase content decreases. This latter effect is illustrated in figure 9.

Silicon and magnesium are rather constant at metal-phase contents greater than 5 per cent. Perhaps the most interesting feature of these two elements is their atomic ratios, which lie close to unity over a wide range of metal-phase concentrations. This is illustrated in table 10.

Concerning the interplay of one element with another, very little can be said in addition to the major effects already discussed. A few points should be mentioned, however, in connection with the prominent oscillations noted in figures 4–8, best exemplified by the oxygen and sulphur curves (figs. 5 and 6). It can be seen quite clearly that when oxygen is a maximum, sulphur is a minimum. A similar effect appears in the case of magnesium (fig. 5), where it will be noted that magnesium maxima and minima coincide with those of oxygen. Combined iron and calcium, on the other hand, appear to possess curves similar to those of sulphur, where the minima coincide with the oxygen maxima. Neither the silicon nor the

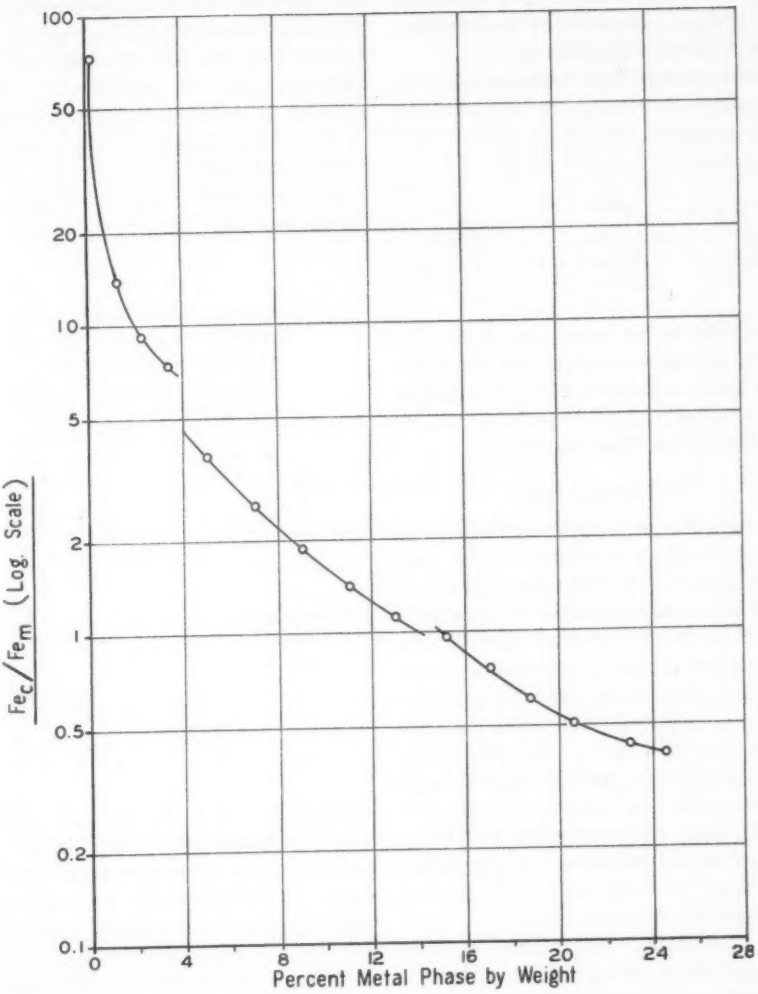

FIG. 9.—The ratio of combined to uncombined iron as a function of metal-phase concentration

aluminum curves appear to bear any relationship (with respect to maxima and minima) to the other curves, although there is some indication of an interplay between these two elements.

Summarizing these minor effects, the composition curves of the elements denoted in figures 4–8 can be divided into three classes:

Class 1	Class 2	Class 3
Oxygen	Sulphur	Silicon
Magnesium	Combined iron	Aluminum
	Calcium	

The curves within each class appear to possess similar structures, and the maxima of class 1 coincide with the minima of class 2. Class 3 appears to be unrelated to either of the other two classes.

INTERPRETATION

Perhaps the most significant feature of the relationships outlined in the preceding section is the observed phenomenon of the rapid increase in the oxidation state of meteorites as their metal-phase contents and total iron contents decrease. It is important that this effect be explained in more concrete terms than the explanation of Prior to the effect that meteorites have separated from a single magma which has passed through successive stages of progressive oxidation. We must inquire into the conditions that might have brought about such marked differences in oxidation states.

Let us suppose, first, that meteorites are fragments of what was once a planet and that the planet was at one time in a liquid state. Let us suppose, further, that all the oxygen associated with this hypothetical planet was combined with the major metallic constituents, notably silicon, magnesium, and iron. Let us further suppose that metallic iron, cobalt, and nickel existed in excess, over and above the amount of oxygen available for combination. Under such circumstances one would expect the liquid to separate into two phases: metallic nickel-iron and magnesium and iron silicates. One would expect to find, in addition, a certain amount of metallic nickel-iron dissolved in the silicate phase. If the planet was small, one would expect the silicate phase to possess a uniform composition throughout and, neglecting surface crystallization effects, the ratio of combined iron to metallic iron to be constant from point to point. If we assume, on the other hand, that the planet was sufficiently large to possess significant gravitational effects, it is reasonable to suppose that a gradient in the dissolved metal would be set up in such a way that the metal content would increase as one approached the center. Unfortunately, however, such a simple picture cannot explain the rapid decrease in combined iron concentration with increasing metal-phase concentration. We must ask whether there is any basis for expecting the combined iron

TABLE 10

THE ATOMIC RATIO OF SILICON TO MAGNESIUM AS A FUNCTION OF METAL-PHASE CONTENT

Metal-Phase Range (Per Cent by Wt.)	Atomic Ratio
0– 1	1.74
1– 2	1.55
2– 3	1.03
3– 4	1.32
4– 6	1.01
6– 8	1.07
8–10	1.12
10–12	1.11
12–14	1.08
14–16	1.15
16–18	1.06
18–20	1.03
20–22	1.05
22–24	0.99
24–26	1.00

concentration to decrease with increasing depth within a planet.

It was demonstrated in the first part of this paper that pressures of the order of 10^5–10^6 atm. can give rise to significant effects in certain equilibria, notably volumes occupied by the various elements within the silicate phase. The large relative volume occupied by oxygen is apparent.

In view of the very small density of oxygen ions, the effect of a large pressure

TABLE 11

IONIC AND ATOMIC RADII OF SOME METEORITIC ELEMENTS

Element	Radius A°	Ref.	Element	Radius A°	Ref.
Fe	1.24	3	Al^{+++}	0.57	1
Fe^{++}	0.75	2	Ca^{++}	1.06	1
Ni	1.24	3	S^{--}	1.74	1
Ni^{++}	0.70	2	Na^{+}	0.98	1
Co	1.25	3	K^{+}	1.33	1
Co^{++}	0.72	2	Ti^{++++}	0.64	1
O^{--}	1.40	2	P^{+++++}	0.34	2
Si^{++++}	0.40	1	Mn^{++}	0.80	2
Mg^{++}	0.78	1	Cr^{+++}	0.64	2

REFERENCES FOR TABLE 11

1. Goldschmidt, 1927, p. 1263.
2. Pauling, 1940, p. 350 (Pauling's correction of Goldschmidt's data).
3. Neuburger, 1936, p. 1.

TABLE 12

RELATIVE VOLUMES OCCUPIED BY ELEMENTS IN METEORITES

Metal-Phase Interval (Per Cent by Weight)	Total Iron	Total Nickel	Total Cobalt	Total Iron, Nickel, Cobalt	Oxygen	Silicon	Magnesium	Aluminum	Calcium	Sulphur	Sodium	Potassium	Titanium	Phosphorous	Manganese	Chromium
Igneous rock	0.46	.....	.....	0.46	94.19	0.75	0.49	0.66	1.28	0.15	1.38	1.86	0.04	....	0.01	0.00
0–1	1.44	0.17	0.02	1.63	90.64	0.64	2.72	0.38	2.29	1.28	0.27	0.09	0.03	....	0.03	0.02
8–10	5.19	0.67	0.06	5.91	83.47	0.56	3.74	0.12	0.56	5.01	0.40	0.16	0.01	....	0.05	0.02
24–26	11.98	0.97	0.07	13.02	75.40	0.52	3.84	0.12	0.59	5.85	0.46	0.14	0.01	....	0.04	0.01

nickel. Similarly, pressure effects might be expected to affect the ratio of combined iron to metallic iron within a planet, if we take into account the large volume occupied by oxygen ions within the silicate phase and the relatively small density of those ions. For comparison table 11 gives the atomic and ionic radii of several elements encountered in meteorites, and table 12 shows the relative

gradient would be to "squeeze" the oxygen toward the outside of the planet. That such a squeezing process could actually break chemical bonds can be seen from the relationship for the chemical potential per gram of a given constituent present in the magma,

$$\frac{d\mu}{M} = (v_x - v_m)\, dP,$$

where v_x is the specific volume of a given constituent (as, for example, oxygen ions), v_m is the specific volume of the magma, and P is the pressure. Within a sphere large enough to give rise to internal pressures of the order of 10^5–10^6 atm., the pressure gradients would be ample to produce shifts in the metallic iron–combined iron equilibrium. The specific volumes of oxygen ions, ferrous ions, and metallic iron atoms are, respectively, 0.4, 0.02, and 0.08 cc/gm. These can be compared with the specific volume of the magma, which is approximately 0.2 cc/gm.

Thus, if iron and iron oxide are to coexist at high pressures and in the presence of a pressure gradient, then necessarily at equilibrium the ratio of combined iron to metallic iron must decrease with increasing pressure. On the basis of such a picture, an increasing combined iron concentration associated with a decreasing metal-phase concentration would be expected. In addition, of course, when the ratio of combined iron to metallic iron becomes so large that the bulk of the iron is in the combined state, the combined iron concentration should pass through a maximum, following which it should decrease with decreasing metal concentration.[6]

Any interpretation of the features of the abundance distribution curves of the other elements encountered in meteorites (silicon, magnesium, calcium, etc.) must necessarily depend upon the detailed chemistry of silicates as a function of temperature and of pressure. However, a few broad features of the curves are reasonably apparent and can be interpreted on the basis of existing data.

The fact that the atomic ratio of silicon to magnesium lies so close to unity for the greater part of the metal-phase range is perhaps significant. The stabilities of magnesium silicates are well known. However, the absolute abundance of magnesium is somewhat less than that of silicon; and, as a result, there is not enough magnesium to combine with silicon on a one-to-one basis throughout. There results an excess of SiO_2, which, being less dense than the magnesium silicates, is squeezed to the outer portions of the planet, together with the silicates of calcium, aluminum, sodium, and potassium.

Crystallization effects, of course, greatly complicate the situation at low metal-phase concentrations. It is noteworthy, however, that magnesium apparently exists on the surface of the earth only through the courtesy of statistical fluctuations. As can be seen from figure 10, the frequency-distribution curve for magnesium in igneous rocks (Richardson and Sneesby, 1922, p. 303) is of a markedly different type from the corresponding distribution curve for magnesium in the silicate phase of stony meteorites. While the curve for meteorites centers around a well-defined maximum, the curve for igneous rock is exponential in character, with the majority of rocks possessing little magnesium. In other words, magnesium has in some manner been excluded either before or during the process of rock crystallization.

A similar correlation holds for other elements common to both igneous rocks and meteorites. In general, if the igneous-rock value for a given element is lower than the value for meteorites of low metal-phase content, the igneous-rock frequency-distribution curve will be exponential in character, with the highest value being at zero per cent. Conversely, if the igneous-rock value is higher than

[6] T. F. W. Barth has discussed the variation of oxygen content with depth in the earth's lithosphere (1948, p. 43).

the value for meteorites of low metal-phase content, the igneous rock frequency-distribution curve will be bell-shaped. The first case is exemplified by magnesium, calcium, and combined iron and the second case by silicon, aluminum, and sodium.

formation is coupled with the chemical relationships existing between igneous rock, plateau basalt, and stony meteorites, the simplest conclusion to be drawn is that meteorites are fragments of a planet similar to the earth in general physicochemical characteristics.

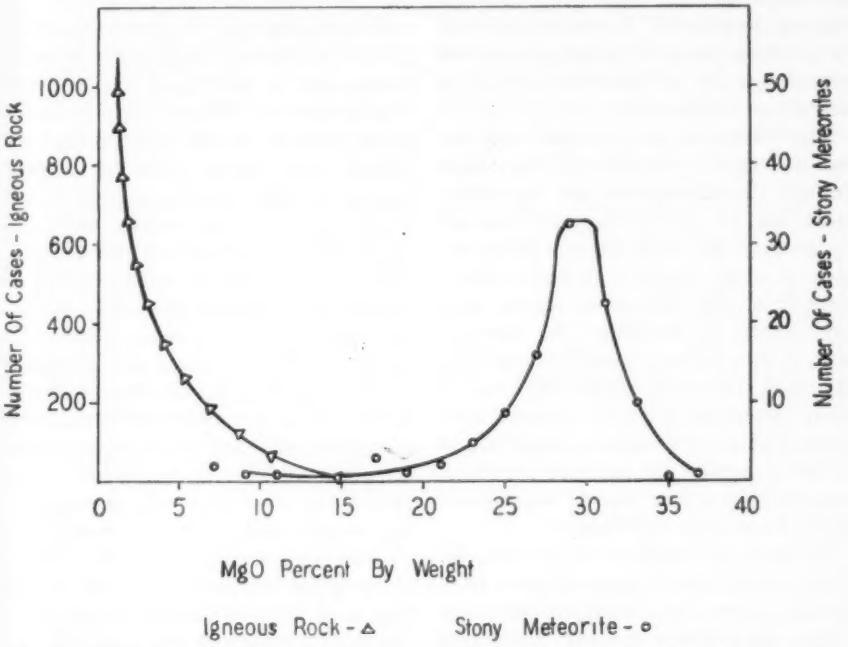

FIG. 10.—Frequency-distribution curves for magnesium oxide in igneous rocks and in the silicate phase of stony meteorites.

PLANET STRUCTURE

We have seen thus far that there exist certain chemical regularities among meteorites involving oxidation-reduction equilibria (or at least something approaching equilibria). It has been demonstrated, furthermore, that the gradations in equilibria from meteorite to meteorite are difficult to explain without assuming the existence of pressure effects upon the equilibria, the order of magnitude of the pressure effects being most probably in the range of 10^5–10^6 atm. When this information is coupled with the chemical relationships existing between igneous rock, plateau basalt, and stony meteorites, the simplest conclusion to be drawn is that meteorites are fragments of a planet similar to the earth in general physicochemical characteristics.

It is possible mentally to construct systems other than a planet which will explain some of the relationships encountered in the sections on phase equilibria in meteorites, and on genetic relationships. However, no system of which the authors have conceived has explained the data nearly so well as the assumption of a body of planetary dimensions. The existing data admittedly are sketchy; and perhaps more intensive study of meteorites in the laboratory will lead to contradictions of the data used in this discussion.

At that time the conclusions drawn here might be changed. However, until that time and to the extent that existing meteoritic data can be trusted, the conclusion appears irrefutable that meteorities at one time were an integral part of a planet. If the conclusion is correct, it is only proper to ask, "What were the properties of the planet?" In order to answer the question, we must investigate the relative frequency of meteorite falls as a function of composition.

The difference in behavior between iron and stony meteorites as they pass through the atmosphere and the differences due to weathering and ease of recognition are well known (Watson, 1939, p. 426). Because of these differences it is most difficult to assess, with any degree of precision, the absolute ratio of iron to stony meteorites striking the earth. However, within the class of stony meteorites it is only reasonable to suppose that such marked differences in behavior do not exist and that a study of frequency as a function of composition might have some significance.

In figure 11 a total of 184 selected meteorite metal-phase analyses[7] have been broken down into numbers of cases versus metal-phase content. Inspection of the figure serves to demonstrate clearly that the frequency distribution of the metallic phase is entirely unlike the distribution of any major constituent thus far encountered in meteorites. Instead of one peak, as is usually the case, the metal-phase frequency curve is spread over a wide range of concentrations, within which three peaks can be rather clearly discerned. The first peak (0–1 per cent metal) and the second peak (7–10 per cent metal) appear definite. Statistical treatment indicates that the third peak (17–19 per cent metal) has a reasonably high probability of being definite. It is noteworthy that the valleys in between the maxima coincide with the discontinuities in the combined iron–metallic iron ratio depicted in figure 9.

If we assume that the meteorites studied are statistically representative of the silicate shell of the disrupted planet and if we assume, in addition, that the metal-phase content of the silicate shell increases with depth below the surface, then it becomes possible to obtain some idea of gross composition as a function of depth. Any precise statement of composition as a function of depth depends, of course, upon detailed knowledge of density as a function of pressure and of the weight of the core of the planet relative to the weight of the silicate mantle. As neither of the quantities can be expressed at present with any degree of exactness, we must confine our discussion for the time being to relative volumes, neglecting compressibility. In this basis, figure 11 has been transformed into a curve showing metal-phase content as a function of "fractional volume increments of the silicate mantle at zero pressure." Inspection of figure 12 serves to demonstrate that the increase in metal content with increasing depth is variable, sometimes being small and at other times being large, and that definite discontinuities exist. It is to be stressed that these discontinuities lie quite close to the discontinuities observed in the combined iron-metallic iron ratios, a curve compiled upon an entirely different basis.

It is, of course, most tempting to connect the metal-phase variations with the well-known seismic discontinuities of second order existing within the earth. If, as the meteorite data indicates, a planet can

[7] Fifty-seven analyses taken from O. C. Farrington's compilation (1911), in which the amount of metal phase had been determined with sufficient accuracy, were added to the 127 analyses comprising table 8.

possess such large metal concentrations and marked gradients of metal content, then one might expect the metal content to play a major role in determining seismic wave velocities. Any attempt, however, to correlate seismic discontinuities and metal-phase contents accurately must await further theoretical treatment of the density of matter under the pressures encountered in bodies of planetary dimensions.

If one assumes that the planet from which meteorites came possessed a core-to-mantle weight ratio comparable to that existing within the earth and if one further assumes that the differentiations within the planet were similar to those now existing within the earth, then certain comparisons can be made: (1) A metal-phase concentration of approximately 0.2 per cent by weight is attained at a depth of 0.02 radial units (130 km.). It is to be noted in this connection that free iron frequently has been observed to exist in certain basalts. (2) The maximum combined iron concentration is reached at a depth of about 0.07 radial units (450 km.). (3) The marked changes in be-

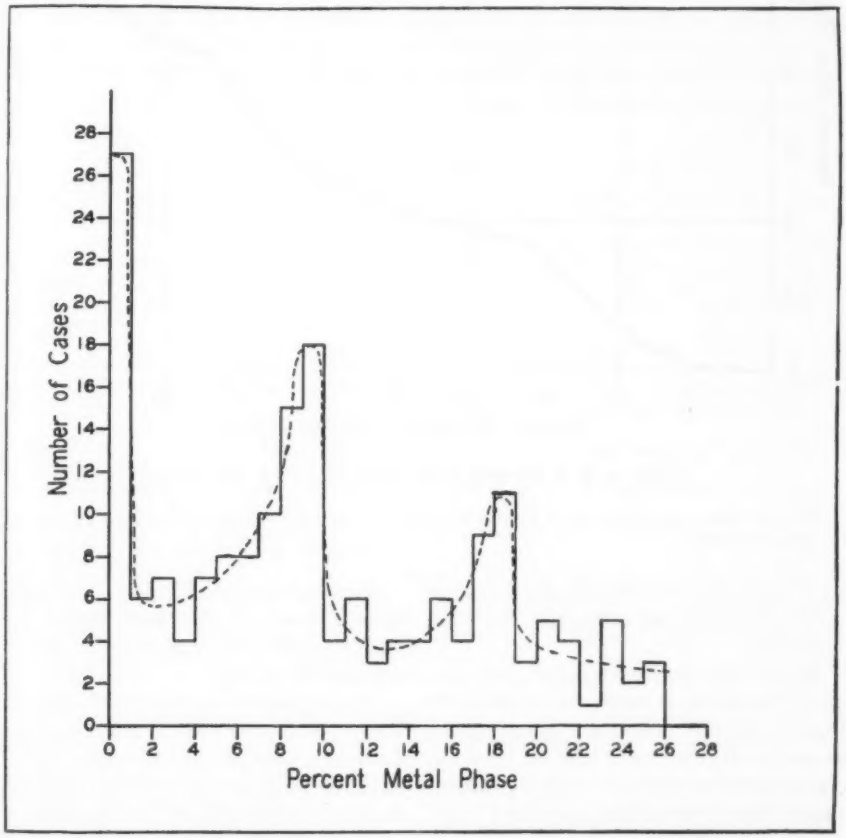

FIG. 11.—Frequency distribution of the metal phase of stony meteorites

havior of magnesium, silicon, calcium, and aluminum occur at a depth of about 0.07 radial units (450 km.). (4) The metal-phase curve discontinuities and the combined iron–metallic iron ratio discontinuities occur at depths of 0.08 radial units and 0.2 radial units (500 km. and 1,300 km.).

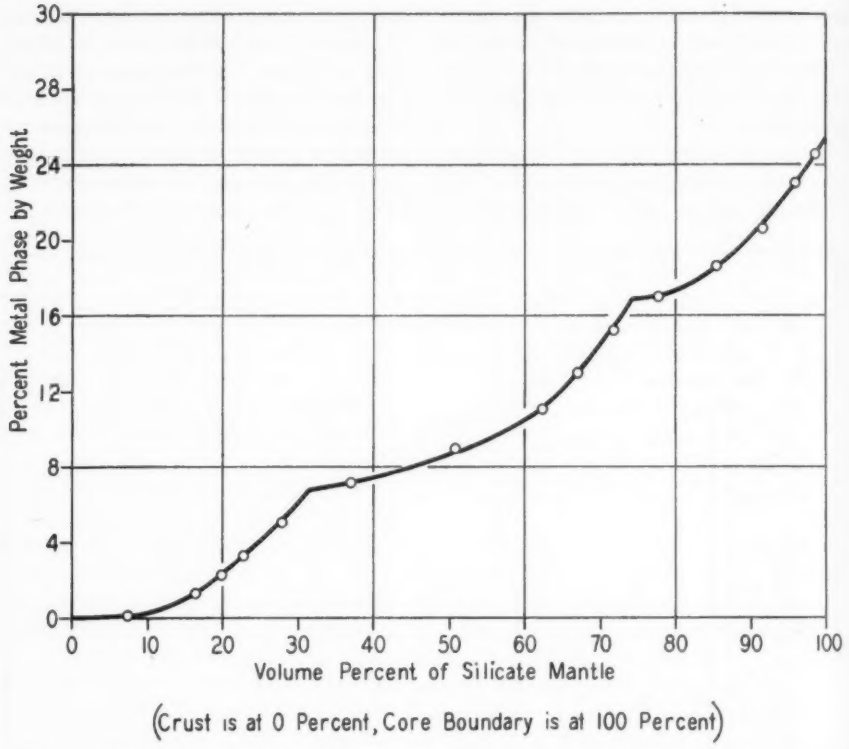

FIG. 12.—Metal-phase content of meteoritic planet as a function of volume percentage increments of the silicate mantle.

Before one takes such figures too seriously, however, it must be emphasized that, if pressure effects are as important as our results lead us to believe, then the chemical differentiations due to pressure should vary from planet to planet (considering only the minor planets) in such a way that the larger the planet, the greater the change in oxidation state with increasing depth. This naturally would result in differences in the relative locations of any discontinuities that might exist.

Summarizing the data concerning planet structure that have arisen from the present study of meteorites, the following can be said:

The meteorite parent-planet (and similarly the earth) was at one time quite hot—in all probability, actually molten. With its composition fixed by the fundamental abundances of the elements involved and the physicochemical processes involved in the initial formation of

the planet, pressure effects produced segregations of matter within the body. The ratio of combined iron to metallic iron for the planet as a whole was fixed by the reducing conditions in existence during the very beginning stages of the formation of the planet. However, owing to the large ionic radius of oxygen (small mass) and to the large density of metallic iron, shifts in the combined iron–metallic iron equilibrium were produced within the body in such a way that the greater the distance from the core, the greater the state of oxidation. As iron was the only major constituent of the magma that could readily give up oxygen, it played a major role in determining the states (combined or metallic) of the minor constituents of the planet. Thus a given metal probably distributed itself between the combined and free states according to the relationship

$$\frac{(M)_{si}}{(M)_{m}} = C\frac{(\mathrm{Fe})_{si}}{(\mathrm{Fe})_{m}},$$

where the ratio of combined iron to metallic iron was determined by the conditions outlined above. The constant C depended upon the prevailing temperature, the thermal properties of the combined and uncombined forms of the minor constituent, the relative volumes of the combined and uncombined constituent, and the pressure. These considerations governed the structure of the planet from the core outward to the point where fractional crystallization effects near the surface produced further changes of substantial magnitude, detailed discussion of which would be beyond the scope of this paper.

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STUDIES FOR STUDENTS

A PREFACE TO THE CLASSIFICATION OF THE SEDIMENTARY ROCKS

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"Perhaps if we stratigraphers insisted on a more refined classification of our sediments, instead of being satisfied with conglomerates, sandstones, shales, limestones, and some minor types, we would make more rapid progress; for it is my belief that precision in classification leads to precision in thought, and so is of vast value as a mental discipline. It might not be amiss to insist that a firm foundation in the classification of our rocks is a needful preliminary to the building of a permanent superstructure, and to urge that we get together and follow the lead of the pyro-petrographers."—A. W. GRABAU (1917).

INTRODUCTION

A great deal of the confusion which pervades present-day thought and usage respecting the textures and structures, the nomenclature, and the classification of sedimentary rocks springs from failure to understand the fundamental character of these deposits and from an inadequate concept of the aim and objectives of rock nomenclature and classification. It seems advisable, therefore, to attempt to analyze and state the problems involved, in the hope that the paths which lead to a theoretically sound and practically useful conclusion will become evident.

This essay, as well as the papers by Shrock and Krynine which follow it, is an attempt to reopen the problem of classification of the sedimentary rocks, in the hope that a sound and workable system may be devised.

BASIS OF ROCK CLASSIFICATION

The first strait jacket in which our thinking has been confined is the time-honored division of all rocks into the three categories: igneous, sedimentary, and metamorphic. This classification has, more than any other one thing, blocked constructive thinking on the classification problem. Grabau (1904) has been almost the only worker to try to break this mental log jam. His effort, though laudable, was bogged down by his attempt to introduce simultaneously an extended and unfamiliar set of rock names. Nonetheless, his recognition of two fundamentally different rock classes, named by him "exogenetic" and "endogenetic," is a sound contribution to the problem. Loosely stated, the exogene rocks are the clastic or detrital rocks; and into this category fall the familiar clastic deposits, such as tuff and sandstone. The endogenetic rocks are, in the vernacular, the "chemical" rocks—precipitates from solution (in the main)—such as rock salt, granite, and the like.

Each of the two main rock groups has its own characteristic textures and structures. Those of a tuff and a sandstone are much more alike than are those of sandstone and rock salt. On the other hand, the textures and structures of granite and rock salt are more akin than are those of tuff and granite. The reasons are self-evident. The clastic rocks have a similar

origin. The principles of fluid mechanics governing the transport and deposition are the same, irrespective of the mineral character of the material deposited. Therefore, the textures and structures of all exogene rocks are much alike. Similarly, the precipitation of materials from solution, governed by phase-rule chemistry, leads to similarity of textures and structures. Differences of temperature and composition of solutions may lead to different mineral composition, but they do not lead to significant differences in texture or structure. The similarity in character of rock salt and granite extends even beyond texture and structure. Invasive or piercement relations are shown by each. This may be more than a coincidence.

Either group, "clastic" or "chemical," may undergo reorganization after deposition. Such reorganization is metamorphism in the broadest sense. Metamorphic recrystallization gives rise to new attributes—the crystalloblastic textures and structures or those induced by recrystallization in the *solid state*. If these changes take place at relatively low temperatures and pressures, they are usually called "diagenetic." Nonetheless, the textures and structures do not differ materially from those induced at higher temperatures or pressures (unless the latter are differential). The differences observed are only those of mineral composition.

GENETIC VERSUS DESCRIPTIVE CLASSIFICATIONS

Much that is wrong has been said about the need for a descriptive classification, independent of a classification based on genesis. There can be no such classification worthy of consideration. Genesis is the ultimate aim of any study of rocks, and no descriptive classification —so called—can be worth much unless the characters used for the classification are meaningful or significant. And the only test for significance is whether they are or are not basic to understanding of origin. Where would the "hard-rock" petrologist be if he were to classify all the coarse-grained igneous rocks on the basis of color? Such terms as "red phanerite," "pink phanerite," "white phanerite," etc., are more or less meaningless. Yet the present-day student of the sedimentary rocks is content to divide the sandstones upon such a chromatic basis.

Genesis must and does permeate our classification. Let one try to describe or define an arkose and distinguish it from a granite without reference to origin. Even the major categories with which most classifications begin—namely, igneous, sedimentary, and metamorphic—are genetic. One must decide origin before one can apply the term "granite" or the term "arkose."

In biology the basis of all sound taxonomic work must be selection of significant characters for classificatory purposes and avoidance of irrelevant peculiarities. Not all organisms with wings, for example, should be grouped together. So, too, any classification of sedimentary deposits, to be usable, must be based on significant properties. It cannot be argued that we do not know what the significant properties are. These are well known. The prevalent confusion, however, stems from failure to recognize the most basic division of rocks into the exogenetic and endogenetic groups, as pointed out above. The significant properties of one group are *not* the significant properties of the other. Hence an attempt to treat all sedimentary rocks alike will and must fail. Thus to apply one set of textural terms to all carbonate rocks obscures, rather than elucidates, their natural history.

THE "HYBRID" CHARACTER OF SEDIMENTS

From what has been said it might be assumed that a rock was either exogenetic or endogenetic. Many are both and therefore exhibit the textures of each. This arises from the fact that such rocks, notably the sandstones (and also certain limestones), contain both clastic elements and a chemically precipitated cement. The textural characteristics are, therefore, hybrid. This hybrid character, however, does not vitiate what has been said above. Instead, it emphasizes the need for recognizing the dual character of many sedimentary rocks. Superimposed on the textures due to mechanical and chemical sedimentation and commonly obscuring them may be those textures induced by recrystallization and replacement (diagenesis). The proper description and interpretation of a sedimentary rock is, therefore, a job of no mean proportions. Sediments may exhibit a textural pattern more complex than that of an "igneous" rock.

Peculiar to the sedimentary rocks and absent from the igneous rocks are the organoform textures and structures. In most cases these are subordinate to the clastic or chemical textures, but in a few rocks they are dominant.

ROCK NAMES

The problem of a proper name for a rock is indeed a difficult one. The students of the sedimentary rocks have attempted to frame a system of nomenclature such that each rock name would be an abbreviated description of the rock (e.g., Grabau). They have also attempted to give a more precise definition of rock terms acquired from the prescientific era. To date, neither effort has been very successful.

One direction in which progress might be made is to follow the example of the igneous petrologist and use a single name, commonly of geographic (type-locality) derivation, for common, recurring rock types. Rather than try to devise names with various prefixes and suffixes, why not use a single term? To the student of igneous rocks each of the terms "rhyolite," "gabbro," or "shonkinite," for example, denotes a rock with a special combination of textures and minerals. Such terms as "arkose" and "graywacke" are comparable terms that are familiar to the student of sedimentary rocks. These can be extended, however, and a term such as "bradfordite" might become a synonym for that particular type of subgraywacke described by Krynine from the Bradford district in Pennsylvania. One commonly refers to "the 'Bedford' (Spergen) type of limestone." Why not "spergenite"?

A QUANTITATIVE CLASSIFICATION

The trend in science is always toward greater precision. So also in petrology. Commendable efforts have been made in recent years to quantify the terms "gravel," "sand," "silt," etc. These have resulted in general acceptance of a few standard size-grades adaptable to clastic sediments. Less successful or complete have been the efforts to extend this quantification to the naming of the aggregate. Though the size-grade "sand" is now established,[1] there is no agreement about the term "sandstone." Must all the material be between $\frac{1}{16}$ and 2 mm.? Must the average fall in this range? What average: median, mode, or mean? And what kind of mean? Or must 50 per cent fall in this range? Or will some

[1] For North American geologists the Udden grades from $\frac{1}{16}$ to 2 mm. are considered sand. European usage is different. So also is the usage of engineers and pedologists.

other proportion be acceptable? No standard usage prevails. But such standards can be established and, no doubt, will be in the future.

TEXTURE AND CLASSIFICATION

The importance of the concepts here outlined can best be made clear by reference to a specific example, namely, a consideration of the textures of limestones. Confusion arises from the fact that limestones do not form a petrographically homogeneous group.[2] They are in part exogenetic or clastic, in part endogenetic ("chemical" and "organic"), and in the main "metamorphic" derivatives of these basic types. Unless this polygenetic character is recognized, no progress can be made toward a description of their textures and structures, or toward a rational nomenclature and classification. To attempt, for example, to impose a single scale of size-grades on such a heterogeneous group of rocks leads only to further confusion. The limestones of clastic origin—which are very common, notwithstanding statements to the contrary—can best be treated as clastics, and the textural terms applied to clastics can readily be applied to these rocks. The Udden geometric grade-scale is appropriate for such deposits. The appropriateness of this (or any other geometric grade-scale) arises from the fact that it tends to symmetrize the size-distribution curves. The tendency of such distributions to be log normal, as shown by this symmetrization, suggests that some underlying law of sorting or grading is operative and that such a scale, therefore, is significant and appropriate. To apply such a scale of size-grades to limestone with chemical textures or to those with "metamorphic" (crystalloblastic textures) is an error. We have no a priori reasons to believe that a geometric grade-scale has any meaning in such cases.

The textures of chemical precipitates ("igneous" textures) are well known, and established practices of description and nomenclature of these textures should be followed. Likewise, the textures of the crystalloblastic rocks, so commonly seen in limestones, should be recognized and described as such.

MINERAL COMPOSITION AND CLASSIFICATION

The mineral composition, like texture, is an important element in any classification scheme. But mineral composition is not wholly independent of texture, and an analysis of the factors affecting the composition must be understood if it is to be rationally used for classificatory purposes.

Minerals are both exogenetic and endogenetic in origin. The chief exogenetic or detrital constituents of a clastic sediment are rock fragments, stable and labile mineral grains from the source rock, and the secondary products of weathering.

The rock fragments of phaneritic or coarse-grained rocks cannot, by the nature of things, occur in fine-grained sediments. Therefore, they characterize only the conglomerates. The fragments of the aphanitic or fine-grained rocks, on the other hand, can occur in the rocks of medium grain size (sands) and do occur in large volume in certain types, notably the graywackes. By and large, however, the rock fragments increase in importance with increase in grain size.

Sand grains are derived principally from the phaneritic igneous rocks. Both stable and labile minerals are present in the sands, but they vary widely in the

[2] The term "limestone," therefore, is, at best, a field term or a term useful in trade or commercial circles.

ratio of one to the other. The monomineralic particles of phaneritic origin dominate in the sands but are present in subordinate amounts in both coarser and finer sediments.

The products of chemical decay are of ultra-fine grain. Hence they characterize the mudstones. Under certain conditions, however, they may be coagulated and be deposited with the sands (as in the graywackes).

In addition to the differences in mineral composition related to the texture of the sediment, there are differences in the mineral composition *in any given size-grade* related to the maturity of the sediment. Maturity may be defined as the measure of the approach of the sediment to the stable end-state. The latter is the ultimate state to which a sediment tends to evolve. A stable sand, for example, would consist almost exclusively of quartz, be exceedingly well sorted, and be highly rounded.

There are various indices of maturity, of which mineral composition is, perhaps, the most important. In the sands, for example, the quartz-feldspar ratio is a very appropriate index. Although this ratio is governed by a number of factors, the prime one is the rate of erosion and concomitant rate of deposition. Inasmuch as the respective rates are controlled by rate of elevation and subsidence, the feldspar content is an index of crustal instability or *tectonism*. Correlative compositional differences due to tectonism may be observed in both the gravel and the clays. The influence of tectonics on the sedimentary regimen has been emphasized in the United States chiefly by Krynine, whose three major families of sediments, designated by their principal arenite member—the orthoquartzite, graywacke, and arkose clans—are indicative of three principal tectonic and geomorphic stages of the geosynclinal cycle.

Superimposed on the compositional differences related to grain size and to tectonism are those due to postdepositional or epigenetic changes—both solution and precipitation. The roles of intrastratal solution and precipitation are just beginning to be understood. The former is responsible for selective loss of the original detrital constituents. The latter is the process to which the mineral cement is due. Least understood are the extensive changes in composition due to simultaneous solution and deposition, that is, the metasomatic replacements.

These epigenetic changes are not, however, wholly independent of the tectonic history of the deposit, as Smithson has pointed out. The sediments of the geosyncline undergo a higher grade of epigenesis than do those of other places. The metasomatic replacement of the limestones and other rocks and the formation of chert seem to be promoted by the rise of the geotherms accompanying geosynclinal downwarp.

Epigenesis is also a function of time. The probability of change is improved with increase in age of the deposit. Therefore, there is a changing lime-magnesia ratio in the limestones with decreasing age, as was early noted by Daly. Similarly, there is an increase in the number and kind of heavy minerals with decrease in age, as noted by several investigators. The ratio of carbonate to silica cement in sandstones of various ages seems also to be a function of time as a consequence of the replacement of the former by the latter.

CONCLUSIONS

Sediments are the product of clastic sedimentation (allogenic), precipitation (biochemical and chemical—authigenic

restricted), and postdepositional solution and precipitation (epigenesis). Each factor modifies the textures, structures, and composition of the sediment in a profound manner and gives rise to distinctive rock types.

A classification scheme, therefore, must be based on parameters which are significant in terms of the origin of the rock under consideration. Inasmuch as the sedimentary rocks are polygenetic, a simple biaxial tabulation of minerals and textures, such as serves for the orthomagmatic igneous rocks, will not do for the sedimentary rocks. Because the mineral composition of the clastic sediments is governed by three factors (grain size, tectonism, and mineral stability after burial), the variations and resultant rock types must be shown by three axes.

The hybrid nature of sediments, expressed by the textural and mineralogical complex which the sedimentary rocks exhibit, the need for greater precision and quantification, and the need for definitive rock names impose further restrictions on one who would propose a classification of the sedimentary rocks.

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A CLASSIFICATION OF SEDIMENTARY ROCKS

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ABSTRACT

A simple field and laboratory classification of sedimentary rocks is proposed. It is based fundamentally on composition and texture, with the primary divisions determined by the mode of origin of the constituents. It is tripartite in nature, consisting of (1) a dominantly fragmental division, including conglomerates, sandstones, and shales; (2) a division represented by rocks which are partly fragmental and partly precipitated, including ironstones, silicastones, coal, limestones, and dolostones; and (3) a division of salinastones, which are dominantly precipitated but may possibly be fragmental. The several new terms in the classification were first proposed in a paper read before the Chicago meeting of the Geological Society of America in December, 1946 (Shrock, p. 1231).

INTRODUCTION

The present classification and taxonomic nomenclature of sedimentary rocks are not altogether satisfactory, for several reasons. Inexperienced students of geology need a simple classification consisting of a few general categories to which they can assign the rock specimens that they collect in the field or store in the laboratory pending detailed study. Certain of these categories have names which have long been used by geologists, e.g., conglomerate, sandstone, shale, ironstone, coal, and limestone. These names are quite satisfactory for our purpose and are used in the proposed classification. Certain other terms, although in use for a long time, are objectionable for one reason or another, e.g., "breccia," "dolomite," "anhydrite," and "gypsum." New terms are proposed to replace these. There is no satisfactory general term for sedimentary rocks consisting largely of silica or for those composed dominantly of the saline minerals. "Silicastone" and "salinastone" are proposed to fill this need. Finally, certain terms have come to have such widely differing or indefinite meanings that they need redefinition, e.g., "graywacke" and "shale." These are discussed and redefined.

The primary purpose of this discussion is to develop a classification of sedimentary rocks simple enough for effective use in the field, where ordinary laboratory and microscope facilities are not available, and in laboratories and museums, where general categories are needed for storage and display purposes. The proposed classification is based primarily on the mode of formation of the particles or crystals—*fragmental* or *precipitated;* secondarily on the two fundamental properties of a sedimentary rock—*composition* and *texture.*[1] These three aspects of sedimentary rocks ordinarily can be determined in the field by observation with the aid of a hand lens and a pocketknife. They can be verified readily in a modern petrographic or sedimentological laboratory.

The complete proposed classification is tabulated on the next page (table 1) and should be consulted in reading the following discussion.

[1] In a prepared discussion to accompany a paper by C. A. Bays and S. H. Folk on "Developments in the application of geophysics to ground-water problems" (1944), P. D. Krynine states: "A rock has really only two basic, fundamental properties; composition and texture, meaning that a rock is made up of certain constituents (generally minerals) put together in a certain way."

TABLE 1

A CLASSIFICATION OF SEDIMENTARY ROCKS

	NATURE OF SEDIMENTS			SEDIMENTARY ROCKS		
DOMINANTLY FRAGMENTAL	Angular particles more than 2 mm. in greatest dimension	Rubble composed of sharpstones		SHARPSTONE	CONGLOMERATE	
	Rounded particles more than 2 mm. in greatest dimension	Gravel composed of roundstones		ROUNDSTONE		
	Angular and rounded particles of rocks and minerals ranging in greatest dimension from 2 mm. to .06 mm.	Volcanic fragments + Tuff Mixture of rock and mineral fragments Quartz + Feldspar Quartz + other minerals in large amount Quartz + other minerals in small amount		TUFFSTONE GRAYWACKE ARKOSE NORMAL QUARTZOSE	SANDSTONE	
	Rock and mineral particles ranging in greatest dimension from .06 mm. to .001 mm. and colloidal particles less than .001 mm. in greatest dimension	Volcanic ash Silt particles — .06 to .001 mm. Clay materials Silt + Clay + Water = Mud		ASHSTONE SILTSTONE CLAYSTONE MUDSTONE	SHALE	
PARTLY FRAGMENTAL	Fe^{II} and Fe^{III} compounds precipitated inorganically and organically as concretions, nodules and layers Impurities commonly present in the layers	Iron concretions Iron compounds + mud, silica, etc.		Concretionary Precipitated	IRONSTONE	PRECIPITATED
	Siliceous inorganic fragments less than .06 mm. in greatest dimension Siliceous organic hard parts and their fragments	Inorganic fragments Diatom frustules, radiolarian skeletons and sponge spicules		Fragmental	SILICASTONE	
	Silica precipitated as oölites, pisolites, etc. Silica precipitated from suspensions and solutions	Silfceous concretions Chert, flint, sinter, etc.		Concretionary Precipitated		
	Plant structures — spores, fronds, leaves, wood, etc. Inorganic sediment Waxes, resins, etc. from decomposition of plants	Plant debris; inorganic impurities Plant fluids		COAL		
	Calcite and Aragonite fragments Calcareous organic hard parts — shells, exoskeletons, plates, spines, and fragments			Fragmental	LIMESTONE	PARTLY
	Organically and inorganically precipitated concretions Inorganically precipitated $CaCO_3$ — Evaporation, etc. Organically precipitated $CaCO_3$ — (1) by NH_3 from decomposition; (2) loss of CO_2 to plants; etc.			Concretionary Precipitated		
	Dolomite fragments Dolomitized organic hard parts			Fragmental	DOLOSTONE	
	Dolomitic concretions			Concretionary		
	Inorganically precipitated dolomite Organically precipitated dolomite			Precipitated		
POSSIBLY FRAGMENTAL	Fragments of anhydrite, gypsum, halite, alkali, nitrate caliche, etc.			Fragmental	SALINASTONE	DOMINANTLY PRECIPITATED
	Evaporites — minerals precipitated during evaporation of saline waters	Anhydrite Gypsum Chlorides Nitrates Other rare salts	Precipitated	ANHYDROCK GYPROCK		

DOMINANTLY FRAGMENTAL SEDIMENTARY ROCKS

INTRODUCTION

Fragmental sedimentary rocks are composed of discrete sedimentary particles, packed together in such manner as to give the mass some coherence. They usually contain more or less interstitial material, which acts as a binder.

The fragments range in size from large blocks many meters in greatest dimension to tiny particles of colloidal size. They vary in shape from irregular sharp-edged fragments, at the one extreme, to spheroidal and ellipsoidal grains, at the other. They may be fragments blown out of volcanoes and related vents[2] or mineral and rock fragments resulting from the disintegration of pre-existing rocks.

In some rocks the interstitial material merely coats the grains around their contacts; in others it partly or completely coats the grains but does not fill all the interstitial voids; in still others it fills the interstices to such an extent that porosity is reduced almost to zero (Krynine, 1941*e*, pp. 108–116). Ordinarily, interstitial material does not enter importantly into the problem of classification. However, if it constitutes greater volume than is necessary for cementation, as in some sandstones, or if it is a major rock constituent, as in coals and shales, it must be given appropriate recognition in defining or describing the rock.

Since, by definition, the rocks of this category are composed of particles or fragments, it follows that the original sediments were granular and could, on deposition, acquire certain features unique to granular materials. The possession of such features (e.g., cross-lamination and ripple-mark) precludes the possibility of the sediments' having been precipitated from solution. This same statement holds for cross-laminated or ripple-marked limestones and silicastones.

CONGLOMERATES

The term "conglomerate" is here used for the rock resulting from a natural binding-together of rubble or gravel. Rubble is any natural accumulation of *angular* mineral, rock, or fossil fragments; gravel, any similar deposit of *rounded* fragments of the same nature (Wentworth, 1935, pp. 225–246). The minimum size of a fragment in these categories is arbitrarily taken as 2 mm. (Wentworth, 1922, pp. 377–392).

Sharpstone conglomerate.—Rubble particles may appropriately be designated "sharpstones." A sedimentary accumulation of such fragments, if bound together strongly enough to form a coherent mass, constitutes a "sharpstone conglomerate." If the interstitial material is as strong as the constituent particles, so that the rock breaks across it and the particles alike, the rock may be considered metamorphosed and designated "sharpstone conglomerite."[3] Acceptance of this nomenclature for sharpstone deposits will obviate further use of

[2] All volcanically ejected materials cooled sufficiently so as not to adhere to adjacent particles on reaching the earth's surface are considered sedimentary. Deposits composed of volcanic particles that are welded together are not considered sedimentary. The reader will find excellent discussions of the several kinds of pyroclastic deposits in two recent papers: (1) C. K. Wentworth and H. Williams (1932, pp. 19–53) and (2) F. G. H. Blyth (1940, pp. 145–156).

[3] B. Willard (1930, p. 438) proposed the term "conglomerite," defining it as follows: "It is suggested, therefore, that the term conglomerate be restricted to those pebbly rocks which break through the matrix and around the pebbles after the manner of sandstones. For the type in which fracture is through the pebbles and matrix, analogous to the conditions observed in quartzite, the term *conglomerite* is proposed."

"breccia," a term which now has several different meanings (Bonney, 1902, pp. 185–206; Norton, 1917, pp. 160–194; Reynolds, 1928, pp. 97–107).

Roundstone conglomerate.—The term "roundstone" was proposed by Fernald (1929, p. 240)

> . . . as a generic term to include the four largest sizes in Wentworth's schedule, boulder, cobble, pebble, and granule [+256 mm., 64–256 mm., 4–64 mm., and 2–4 mm., respectively] . . . [and] . . . to designate the unassorted accumulations composed of two or more sizes of rounded stones that occur in many situations.

Roundstones, therefore, can be thought of as the components of gravel, in the same way that sharpstones are the components of rubble. It is suggested that the rock composed of naturally cemented roundstone gravel be designated "roundstone conglomerate," and its metamorphosed equivalent, "roundstone conglomerite."

Types of conglomerates.—The student has only to observe the shape of the particles of a conglomerate to determine which of the two types it is. He can then use noun and adjectival modifiers to define the rock more exactly. Qualifying terms are available for expressing almost any aspect of the rock. Thus there are chert sharpstone conglomerates, quartzite roundstone conglomerates, and conglomerates described as basal, edgewise, desiccation, intraformational, coralline, shell, etc. Many other types have been described (Grabau, 1904, pp. 228–247; Mansfield, 1907, pp. 550–555; Field, 1916, pp. 29–66; Barrell, 1925, pp. 279–342; Twenhofel, 1936*b*, pp. 677–703; Allen, 1936, pp. 18–47; Pettijohn, 1943*b*, pp. 387–397), and the student will usually find, upon consulting the literature, that some geologist has previously described a conglomerate similar to the one he is studying.

SANDSTONES

The term "sand" has long been used for any sedimentary accumulation of discrete mineral, rock, and fossil particles of a certain size—2–0.06 mm. in greatest dimension—regardless of their composition or mode of origin. Thus there are olivine, magnetite, glauconite, and quartz sands, for example; there are also volcanic sands, oölitic sands, coral sands, and foraminiferal sands. Many others have been enumerated (Allen, 1936, pp. 18–47; and Smith, 1946, pp. 121–143). Suffice it to emphasize that the term "sand" has a broad connotation rather than being limited to specific kinds of particles.

It would be logical to define a sandstone as a rock composed of sand, but such a broad definition would not be acceptable to most geologists (e.g., a rock composed of foraminiferal shells or calcareous hard parts of other animals would be called a "limestone," even though the particles of which it is composed would have been designated "sand" when they lay on the beach or on a shallow bottom). Therefore, "sandstone" is here used in a somewhat restricted sense for a sedimentary rock composed dominantly of quartz grains, feldspar grains, or bits of rock or mixtures of these, with minor amounts of other minerals.

For the present purpose five types of sandstone are included in the proposed classification. These are designated by second-order terms: "tuffstone," "graywacke," "arkose," "normal sandstone," and "quartzose sandstone." As with other sedimentary rocks, noun and adjectival modifiers can be used when necessary to define the nature of a sandstone more fully (e.g., glauconitic sandstone and micaceous sandstone). The reader will find an excellent discussion of sand-

stones in the previously cited article by Allen (1936, pp. 18–47).

Tuffstone.—Sedimentary accumulations of volcanic fragments of sand size might well be designated "tuffstone" and considered a special type of sandstone. There are many tuffstones of this nature in the geologic section, especially in the Pre-Cambrian, and they deserve a specific rock name (Bailey, 1926; Wentworth and Williams, 1932, pp. 19–53, Blyth, 1938, pp. 392–404; 1940, pp. 145–156).

Graywacke.—The term "graywacke," which has been much misused (Krynine, 1941*a*, pp. 2071–2074), may be simply defined as *a sandstone composed of a mixture of rock and mineral fragments ranging in greatest dimension from 2 to 0.06 mm.* This definition conforms with the classical meaning of the term, as well as with preferred modern usage. Knopf (in Pirsson and Knopf, 1926, p. 340), for example, defines the term as follows:

> *Graywackes.*—These are sandstone-like rocks of a prevailing gray color, sometimes brown to blackish, which, in addition to quartz and feldspar of an arkose, contain rounded or angular bits of other rocks, such as fragments of shale, slate, quartzite, granite, felsite, basalt, etc., or of varied minerals, hornblende, garnet, tourmaline, etc.

This definition, as well as the simpler one stated earlier, satisfactorily describes the Third Bradford sand of Pennsylvania, which Krynine (1940) considers a typical graywacke.

Graywackes are of great importance the world over in the earlier pre-Cambrian sedimentary record, sharing with arkoses a dominance over normal sandstones (Pettijohn, 1943*a*, pp. 925–972). If the term "graywacke" is to have significance with reference to geosynclinal deposition—and this is highly desirable (Jones, 1938; Krynine, 1941*c*, p. 1916)—then it should be used in its classical sense.

Arkose.—An arkose is a sandstone containing a relatively high percentage of feldspar grains (i.e., 25–50 per cent by weight). Many sandstones contain a few scattered particles of feldspar; arkoses, however, are composed dominantly of the two minerals quartz and feldspar. The term "arkose," as used by many geologists, also has genetic significance (Krynine, 1941*b*, pp. 1918–1919).

Normal sandstone.—The term "normal" is applied to those sandstones composed dominantly of quartz but with large amounts of other minerals. Normal sandstones are widely represented in the geologic section from the oldest to the youngest rocks. They exhibit considerable variation in the nonquartz components and in some cases contain small amounts of rock particles in addition to the mineral grains. They lie between graywackes, on the one hand, and quartzose sandstones, on the other. The typical normal sandstone does not have a conspicuous amount of any mineral other than quartz; some, however, are characterized by large amounts of nonquartz minerals (e.g., the glauconitic sandstones of the Wisconsin Upper Cambrian [Twenhofel, 1936*a*, pp. 472–487] and of the New Jersey Cretaceous). In the latter case the appropriate noun or adjectival modifier can be used to indicate the abundant nonquartz mineral (e.g., glauconite or glauconitic sandstone and micaceous sandstone).

Quartzose sandstone.—The term "quartzose" is applied to those sandstones composed almost entirely of quartz grains. The nonquartz minerals are typically small in amount (generally less than 1 per cent) and restricted to a few durable species (e.g., zircon, tourmaline, garnet, and ilmenite). Typical representatives of this type of sandstone are the St. Peter and Oriskany. Quartz-

ose sandstones are composed of multicycle sand grains representing sediments that have been eroded and deposited repeatedly. As a consequence, the grains are frosted in some rocks and are commonly well rounded.

Metamorphosed sandstones.—Silica-bound quartzose sandstones in which fracture crosses interstitial silica and quartz grains alike are designated "quartzites." Several kinds of quartzites are now generally recognized (Krynine, 1941*d*, pp. 1915–1916). "Arkosite" (Gruner, 1941, pp. 1577–1642; and Pettijohn, 1943*a*, pp. 925–972) has been applied to metamorphosed arkoses of the Lake Superior pre-Cambrian; and "graywackite," if it has not already been proposed, might well be used for metamorphosed graywackes.

SHALES

Shale is a laminated argillaceous rock, composed of a complex of silt fragments, colloidal micelles (which consist in a dispersed medium of a particle ranging in maximum dimension from 1 μ [0.001 mm.] to 1 mμ [0.000,001 mm.], adsorbed ions, and bonded water), and impurities such as organic materials. The rock is sufficiently consolidated and lithified that it maintains its essential character when weathered (especially when wetted after having been dried out).[4] Inasmuch as shales vary greatly in degree of consolidation and lithification, in the relative amounts of constituent materials, and in the development of lamination, to say nothing of their content of sand grains and organic matter, it seems advisable, and, in fact, it has now become rather common practice, to recognize by name three distinct types of subshale rocks—*siltstone*, *claystone*, and *mudstone* (Twenhofel, 1937, pp. 81–106). In addition, sedimentary deposits of volcanic ash deserve a special designation, and the term "ashstone" is suggested. Although "colloidstone" has been proposed (Alling, 1943, p. 266) for sedimentary rock composed of particles smaller than 1 μ (0.001 mm.), that term is not used in the proposed classification because the size range that it covers is included in the definition of shale.

Ashstone.—The geologic section contains many sedimentary deposits of volcanic ash. Such rocks are important and distinctive enough to be given a special designation. "Ashstone" is suggested for rock composed of particles of volcanic ash less than 60 μ (0.06 mm.) in greatest dimension. (Some writers would prefer a smaller figure.) Several terms have already been proposed for the finer-grained pyroclastics, but these apply more appropriately to recent accumulations (Wentworth and Williams, 1932, pp. 19–53; and Blyth, 1938, pp. 392–404; 1940, pp. 145–156).

Siltstone.—The term "siltstone" is in common use for indurated silt (Wentworth, 1922, p. 381). It is composed chiefly of mineral, rock, and fossil fragments with maximum dimensions from 0.06 ($\frac{1}{16}$ mm.) to 0.0039 mm. ($\frac{1}{256}$ mm.). The definition of siltstone given above follows general usage with respect to the limiting dimensions of the constituent particles (i.e., 0.06–0.0039 mm.). The author, however, prefers to extend the smaller dimension to 1 μ (0.001 mm.) in order to include particles now commonly referred to the so-called "clay size." Inasmuch as particles between 0.004 and 0.001 mm. in greatest dimension can be of some mineral other than a clay mineral (e.g., quartz), it would seem advisable to eliminate the term "clay" as a size

[4] It is pointed out later that mudstone, which may be considered a type of subshale, slakes to mud when dried out and wetted repeatedly.

designation. Furthermore, since 1 μ is the wave length for visible light and has been taken by several investigators as the upward limiting dimension for a colloidal particle, it is a logical dimension to use as a downward limit for silt particles. Therefore, accepting the suggested modification of the definition of silt not only would obviate further use of clay as a size term but would also fix the downward limit of a silt particle at the beginning of the colloidal range. This modification is incorporated in table 1. Siltstone feels harsh to the touch and does not lose its coherence when dried out and wetted repeatedly.

Claystone.—"Claystone" has had wide usage for very fine-grained, somewhat unctuous, conchoidally fracturing sedimentary rock composed largely of clay material (Grim, 1942, pp. 225–275). Claystone spalls when weathered, but it does not lose its solidity, i.e., it does not become mud.

Mudstone.—"Mudstone" is an appropriate term for those partly indurated argillaceous rocks which slake readily to mud when dried out and wetted repeatedly. The term has been in use for over a century, but not always with the connotation just given (Twenhofel, 1937, p. 90).

Designation and description of shales.—Ashstone, siltstone, claystone, and mudstone may all be regarded as partly lithified shales. Therefore, if they cannot be identified certainly in the field, they can be designated "shale" until their exact nature can be established in the laboratory.

All sorts of modifiers are available for indicating the many characteristics of argillaceous rocks. A few examples are ferruginous claystone, carbonaceous mudstone or shale, black shale, sandy (=quartz) siltstone, pencil and paper shales, oil shale, and calcareous shale.

Metamorphosed shales.—Argillite, slate, phyllite, and schist are members of a series of metamorphosed argillaceous materials. The sedimentary origin of argillites and slates is made evident by such features as graded bedding and coarse sandy layers. Phyllites and schists, on the other hand, rarely possess inherent features by which it can be certainly established that they were derived from sedimentary rocks.

TILLSTONE

The few deposits of ancient glacial till that have been recognized as such are usually referred to as "tillite." It would make for more uniform classification if "tillstone" were adopted for consolidated and "lithified till" and "tillite" were reserved for metamorphosed tillstones. This modification of terminology should not cause too much disturbance, inasmuch as some of the better-known ancient till deposits (e.g., Gowganda, Dwyka, and Squantum) are more or less metamorphosed, hence are tillites (Coleman, 1926). "Tillstone" is not included in the proposed classification because it implies genesis. A tillstone would fall in one or the other of the two categories of conglomerate.

PARTLY FRAGMENTAL AND PARTLY PRECIPITATED ROCKS

INTRODUCTION

The rocks included in this indefinite subdivision are alike in being composed of mixtures of organic and inorganic fragments or concretions and precipitated compounds.

The particles may be fragments resulting from the breaking-up of organically secreted hard parts or structures (e.g., broken and comminuted corals);

they may be skeletal or shell fragments or the skeletons and shells themselves, released upon decomposition of enclosing organic tissue (e.g., sponge spicules, echinoid plates, radiolarian skeletons, foraminiferal tests, and diatom frustules); or they may be concretions (e.g., oölites and pisolites), built in several different ways. They may, on the other hand, come from an inorganic source (e.g., siliceous inorganic materials resulting from the weathering of chert and cherty limestones and included under the term "tripoli" [Metcalf, 1946, pp. 1–25]).

The precipitated parts of the rocks were produced by organically or inorganically induced flocculation of colloids and precipitation of soluble salts. These substances usually form the matrix of the rock, filling the voids between fragments and concretions (e.g., radiolarian silicastone and oölitic silicastone). They constitute the entire rock in some limestones and dolostones and in many silicastones.

Suffice it to point out that the five types of sedimentary rocks included in this general category show considerable range in composition and texture and in the proportions of fragmental and precipitated constituents. It will be necessary, therefore, as in the previous types, to employ descriptive modifiers to indicate the special characteristics of any general rock type.

IRONSTONES

The term "ironstone" has commonly been applied to thin beds of hard, tough, iron-bearing argillaceous rock that is characteristic of coal-bearing sequences the world over. It might also be applied to other sedimentary rocks in which compounds of iron are a major constituent (e.g., a lateritic iron ore, such as that lying between Miocene lava flows in Oregon [Williams and Parks, 1923, pp. 1–44]). Oölitic iron ores of the Clinton type represent concretionary ironstones.

SILICASTONES

"Chert," "flint," "hornstone," and "novaculite" are familiar terms for certain siliceous rocks. "Diatomite" and "radiolarite" have been applied to rocks composed of the siliceous frustules of diatoms and the siliceous skeletons of radiolaria, respectively. There are certain siliceous rocks composed largely of concretions (mainly oölites) and others, which are rare, that are made up of very fine silica fragments (tripoli) derived originally from the chemical weathering of chert, cherty limestones, and novaculite. There is no satisfactory general term for this group of widely different, typically fine-grained siliceous rocks. "Silicastone" is suggested as such a general term. If it is desirable, the rock can always be classified more precisely, after necessary laboratory work, by using appropriate modifiers (e.g., cherty silicastone, diatomaceous silicastone, and oölitic or concretionary silicastone). In the field the more general term "silicastone" can be used when there is doubt concerning the true nature of the siliceous rock.

Much has been written about certain silicastones, and some controversy still exists over the origin of a few of them (Tarr, 1926, pp. 1–46; 1938, pp. 8–27; Tarr and Twenhofel, 1932, pp. 519–546).

COALS

Coal is a complex of plant debris, solidified organic compounds derived from plant tissue, and inorganic impurities washed or blown into the coal swamps from the surrounding land areas (White and Thiessen, 1913, pp. 1–390;

Thiessen, 1947, pp. 1–53; Stutzer and Noé, 1940, pp. 1–461; Moore, 1940). It is unnecessary for the present purpose to discuss the several kinds or grades of coal. The exact nature of a coal can be determined only in a well-equipped laboratory. The beginning field geologist needs only the general term for his notes.

The proposed classification does not include deposits of solid hydrocarbons, such as asphalt, gilsonite, etc. These represent special types of sedimentary deposits and are exceedingly rare as compared with those included in the classification.

LIMESTONES

The term "limestone," as it is now used almost universally in North America, includes three distinct kinds of calcareous rocks:

1. *Fragmental*, consisting of the fragments of calcareous fossils, of calcite crystals, and of broken concretions
2. *Concretionary*, composed of oölites, pisolites, stromatolites, biostromes, bioherms, and other masses of calcareous material built by organisms or formed inorganically
3. *Precipitated*, consisting of fine-grained calcareous material precipitated from solution either organically or inorganically

Fragmental limestones, the most abundant of all calcareous rocks, commonly exhibit typical sedimentary features of granular deposits, such as ripple-mark and cross-lamination. Concretionary limestones are likely to be local in extent, variable in thickness, and uneven in texture. Precipitated limestones, which are rare as compared to the two preceding types, are likely to be very fine grained, uniformly textured, commonly somewhat argillaceous, and in some cases mud-cracked (Tarr, 1925, pp. 252–264; Gee, 1932, pp. 162–166; Wood, 1941, pp. 192–200). Precipitated limestones lack all sedimentary structural features which depend upon a megascopic granular condition of the original sediments.

It is considered more desirable to use the term "limestone" in its broad sense than to propose new names for the three types just described. As in previous types, modifiers can be used to indicate special characteristics of the rock. A few such words are "chalk" and "marl" ("chalkstone" and "marlstone" might be appropriate rock names in certain cases), "travertine," "tufa," "coquina," "shell," "coralline," "algal," "dolomitic," "hydraulic," "lithographic," and "biohermal" or "reef."

Marbles, the metamorphosed equivalents of limestones (and dolostones), are of great variety because the calcareous rocks from which they were formed were widely different. It is worth while to emphasize that fragmental marbles come only from fragmental limestones. As with other metamorphosed sedimentary rocks, some marbles have been so completely recrystallized that any original sedimentary textures and structural features have been totally destroyed, making uncertain the nature of the original calcareous rock.

Certain limestones are recrystallized early in their history and have some of the properties of metamorphic marbles. These have been designated "diagenetic marbles."

DOLOSTONES

The term "dolostone" is proposed for those sedimentary rocks consisting largely of the mineral dolomite, such as the Niagaran dolomitic rocks of Illinois (Willman, 1943, pp. 1–89). Adoption of this term would avoid the confusion arising from the use of a mineral name for both a mineral species and a rock type.

The presence in dolostones of other minerals and substances and of certain textural and structural features can be indicated by appropriate modifiers (e.g., calcitic, oölitic, and saccharoidal dolostone). Many rocks which have been described as dolomites might better be called "dolomitic" or "magnesian limestone" if dolomite is not the dominant mineral.

Dolomite marbles resulting from metamorphism of dolostones seem to be rare. Recrystallized dolostones, on the contrary, are not uncommon.

DOMINANTLY PRECIPITATED AND POSSIBLY FRAGMENTAL ROCKS

This small subdivision includes those saline rocks usually called "evaporites." "Salinastone" is suggested as a general term for this group of sedimentary rocks.

The commonest salinastones are those composed of gypsum and anhydrite. "Gyprock" has already found use for the former, and "anhydrock"[5] is suggested as a suitable term for the latter. The composition of other less common salinastones can be indicated by appropriate modifiers (e.g., halite salinastone and borate salinastone).

SUMMARY

The classification of sedimentary rocks here suggested is based primarily on whether the constituents are fragmental or precipitated and secondarily on composition and texture. Both sets of properties can usually be determined readily in the field by direct observation and by simple tests. Texture and composition are the basis of the general group terms (sharpstone and roundstone conglomerates, sandstone, shale, ironstone, silicastone, coal, limestone, and dolostone), which are names of the first order. Distinct subtypes (e.g., tuffstone, graywacke, arkose, ashstone, siltstone, claystone, and mudstone) bear names of the second order. Noun and adjectival modifiers constitute terms of the third order. It should be possible to assign a first-order name to any common sedimentary rock encountered in the field. Second- and third-order names had best be reserved for use in the laboratory after the nature of the rock has been determined in greater detail.

Students will find the several categories of sedimentary rocks of the proposed classification fully described in such standard works as *Treatise on Sedimentation* (Twenhofel, *et al.*, 1932); *Sedimentary Petrography* (Milner, 1940); *A Handbook of Rocks* (Kemp, 1940); and *Rocks and Rock Minerals* (Pirsson and Knopf, 1947).

Recent studies of oil sands (Krynine, 1940), microlithologies (Alling, 1945, pp. 737–755), and textures of carbonate rocks (De Ford, 1946, pp. 1921–1928), to cite but three, are good indices of the rich field of research that lies ahead for the beginning student of sedimentary rocks. It is hoped that the classification here proposed will serve the student as a simple filing system, which he can elaborate and refine as the need arises.

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[5] "Anhydrock" was suggested to the writer by his colleague, Professor W. L. Whitehead, of Massachusetts Institute of Technology. It seems to be the most euphonious word that can be formed from "anhydrite" and "rock."

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THE MEGASCOPIC STUDY AND FIELD CLASSIFICATION OF SEDIMENTARY ROCKS

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PART 1. DEFINITION OF BASIC TERMS AND FUNDAMENTALS OF COMPOSITION, TEXTURE, AND STRUCTURE

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PART 1. DEFINITION OF BASIC TERMS AND FUNDAMENTALS OF COMPOSITION, TEXTURE, AND STRUCTURE

The necessary knowledge is that of what to observe.—EDGAR ALLAN POE.

INTRODUCTION

The systems of classification of sedimentary rocks currently in vogue are either entirely too generalized or tend to combine several dissimilar features, generally texture (and, specifically, grade size), together with some other factor, such as bulk chemical composition or mode of genesis, which, regardless of their possible combinations, cannot result in an unequivocal classification. In-

deed, one of these factors may be either somewhat subjective (genesis) or at least not susceptible of objective and accurate description at the megascopic level (bulk composition). As a result, it is extremely difficult to translate some of the existing sedimentary names into quantitative, objective, and reproducible terms; and, conversely, many of the same names when employed by different workers may carry entirely different meanings.

At present the basis of sedimentary classification is somewhat arbitrary, with four main systems in general use. These classification systems, which may be combined in several ways, are as follows:

1. By mode of origin (very common)
   - Clastic or fragmental rocks
   - Chemical precipitates
   - Biogenic products (plant and animal)
2. By medium in which rocks originate
   - Aqueous (water-laid)
   - Eolian (wind-laid)
   - Glacial (ice-laid)
3. By bulk composition
   - Arenaceous (sandy)
   - Argillaceous (clayey)
   - Calcareous (limy)—although many other chemical precipitates are frequently added to this scheme
4. By texture and grain size, with many variations, permutations, and the addition of much involved terminology
   - Psephites (coarse clastic rocks)
   - Psammites (medium clastic rocks)
   - Pelites (fine clastic rocks)
   - Crystalline (chemical rocks)
     - Phanerocrystalline (coarse and medium)
     - Aphanitic (fine and very fine)

None of these schemes can compare in completeness, simplicity, or objectivity with the mineral composition–texture scheme in use for the classification of igneous rocks.

As a result, considerable scientific confusion is found to prevail in published descriptions of sediments. Attempts by the writer to teach (mostly unsuccessfully) undergraduate students these different unco-ordinated schemes of sedimentary classification, after putting them through the simple and logical nomenclature of igneous petrography, showed only too well how such confusion is likely to develop.

This confusion is undesirable not only from the point of view of "pure science" but also from the applied, economic aspect. Indeed, it should be clear to everybody that, in the field of igneous, so-called "hard-rock" mining geology, an economic geologist who is not capable of differentiating a granite from a gabbro and who insists in lumping these two rocks together under the vague term of "phanerite" is not exactly an asset to any employer. In fact, such a man would not be able to hold his job for one single day.

However, in the current practice of petroleum geology all the medium-grained clastic rocks are usually lumped together under the term of "sandstone." It happens, however, that a successful search for stratigraphic traps requires, as a necessary working tool, a much more precise and refined terminology than such a vague and ill-defined catch-all term.

As a result of work on the geosynclinal sediments of the Appalachian region from 1937 on, the writer finds that there appears to be a series of relationships between the genesis, composition, structure, and texture of sediments. Some of these concepts were presented in preliminary form elsewhere (1942, pp. 537–561; 1943; 1945, pp. 12–22) and others will be published in a complete form in the near future.

In the meantime this work has led to the formulation of a simplified form of sedimentary rock classification based on mineral composition and texture which forms the subject of the present discussion. This classification has been used

with success at the Pennsylvania State College since 1942 for teaching advanced sedimentology to Seniors and graduate students and since 1944 for teaching elementary hand-specimen petrology to Sophomores.

The research work that led to the formulation of these concepts was financed under Research Projects E-6 and E-14 by the Mineral Industries Experiment Station of the Pennsylvania State College, Dr. A. W. Gauger, director. To his colleague, Dr. J. C. Griffiths, the author is indebted for several excellent suggestions that have been incorporated into this paper.

THE MAKEUP OF SEDIMENTS

A rock, or any other solid for that matter, has only two basic, fundamental properties: composition and texture, meaning that a rock is made up of certain constituents (generally minerals) put together in a certain way. All other properties, such as color, density, and similar mass properties are only derived properties. Even structure is not entirely a primary property but is rather the reflection of changes—abrupt or gradual, horizontal or vertical—in texture and composition within one formation or between different formations. However, for practical purposes, structure can be considered as the third major property of sediments when describing entire sedimentary bodies.

A tenable (that is, objective and reproducible) classification of sediments must be based for hand specimens on composition and texture and for entire sedimentary bodies on composition, texture, and structure, with possible qualifications introduced by the addition of some of the subelements of texture (size, homogeneity) and possibly some of the principal derived properties, such as color.

There are altogether ninety-three important derived properties or parameters of sedimentary rocks, but most of these are quite unnecessary for purposes of megascopic identification.

THE COMPOSITION OF SEDIMENTS

MAJOR CONSTITUENTS

Just as it is impossible to understand the properties and probable behavior of an engineering structure without knowing of what materials it has been built, so it is equally impossible to understand a sedimentary rock without knowing what it is made of, and by this is meant knowing rather than guessing. Hence an adequate working knowledge of elementary mineralogy is an absolute prerequisite for the field study of sediments.

Although over one hundred and sixty different minerals have so far been identified in sediments, less than twenty mineral species form well over 99 per cent of the bulk of sedimentary rocks. It is rare indeed that more than five or six minerals occur in sizable amounts in any one rock. Thus the exact determination of the composition of a sediment (shales excepted) is not a particularly difficult affair.

The twenty minerals which form the bulk of the sedimentary rocks are known as "major" or "main" constituents. A major constituent is defined as one that forms at least 1 per cent of a rock. A further subdivision of major constituents into more abundant varieties, which form over 10 per cent of some common rocks, and less important ones, which form only 1–10 per cent, is shown in table 1.

In addition, approximately twenty other minerals, known as "accessory minerals," occur in very small amounts in sediments, although locally they may be of great importance. These accessory minerals, however, are difficult to study megascopically. A list of the major con-

stituents and of the more important accessories is given in table 1.

An understanding of sediments may be condensed to an understanding of the relationships which exist among the

These minerals are not distributed in sedimentary rocks in a haphazard manner but tend to occur in certain definite associations, which form a series of petrographic end-members.

TABLE 1*

THE COMMON MINERALS OF SEDIMENTS

	Major Constituents		Accessory Minerals (Less than 1 Per Cent of Rock)
	Over 10 Per Cent of Rock	Less than 10 Per Cent of Rock	
Detrital Minerals	QUARTZ Microcline CLAY MINERALS (kaolin-bauxite) FINE-GRAINED MICAS (illite, sericite muscovite)	DETRITAL CHERT Sodic plagioclase (albite-oligoclase) Coarse-grained micas: muscovite biotite chlorite Hematite Limonite	"IRON ORES": MAGNETITE, ilmenite, DETRITAL LEUCOXENE STABLE GROUP: ZIRCON TOURMALINE, rutile UNSTABLE GROUP: APATITE, EPIDOTE GARNET, HORNBLENDE kyanite, sillimanite staurolite, titanite zoisite MICAS: frequently occur as accessories rather than as major constituents
Chemical and Authigenic Minerals	CALCITE DOLOMITE ANKERITE	CHERT and opal "SECONDARY" QUARTZ GYPSUM and anhydrite, halite Some hydromicas of the illite-sericite-chlorite series Phosphates and glauconite, Siderite and some iron ores	ANATASE; authigenic rutile and leucoxene

* Minerals printed in capitals are the relatively more common ones within each group.

major constituents. Most of these relationships are established at the very beginning of the "life" of a sediment and generally are a function of its mode of formation.

Computations by the writer have shown that the average sediment consists of the following minerals and mineral families:

	Per Cent
Quartz	31.5
Chalcedony (chert)	9.0
Feldspars	7.5
Micas and chlorite	19.0
Clay minerals	7.5
Carbonates	20.0
Iron oxides	3.0
All others	3.0

PETROGRAPHIC END-MEMBERS

The composition of *all* sedimentary rocks can be reduced to two basic groups of end-members, which may be mixed in all proportions in a purely mechanical way.

1. A detrital fraction consisting of solid material brought in as *solid detritus* from outside the basin of deposition and precipitated through settling within this basin. In 99 per cent of the cases this detrital fraction is made up of silicates. The composition of this detrital fraction depends on the petrology of the source areas and the intensity (and effectiveness) of chemical weathering and erosion

within the source area, plus some modification during transport.

2. A chemical fraction existing as a *solution* within the basin of deposition and precipitated chemically. This chemical precipitation may proceed through inorganic or biogenic agencies. The chemically precipitated material may not move on the sea floor after its precipitation, and then it develops a crystalline texture. Or it may be shifted about by bottom currents (particularly true of organically precipitated rocks composed of shell fragments), and then, although its origin is chemical, its texture will be clastic. More than three-quarters of this chemical fraction is made up of carbonates, and most of the balance of silica. Other constituents (glauconite, phosphates, iron oxides) may be very abundant locally but are relatively rare on a volumetric grand-total basis.

The fundamental difference between sediments and igneous rocks is that sediments consist of *two* major groups of end-members (detrital and chemical), whereas igneous rocks are made up of only one (generally chemical, tuffs excepted). The only possible direct analogy of a sediment within the igneous group would be a rock formed from the consolidation of, let us say, a basaltic lava flow into which fell a shower of rhyolitic volcanic ash. The basalt would be the chemical end-member (from solution), the rhyolitic ash the detrital (solid) end-member. Such an igneous rock, although theoretically possible, would be a very rare freak indeed. But every aqueous sediment, without a single exception, has been formed exactly in this way, by a mixing of solid detrital and chemically precipitated material (fig. 1).

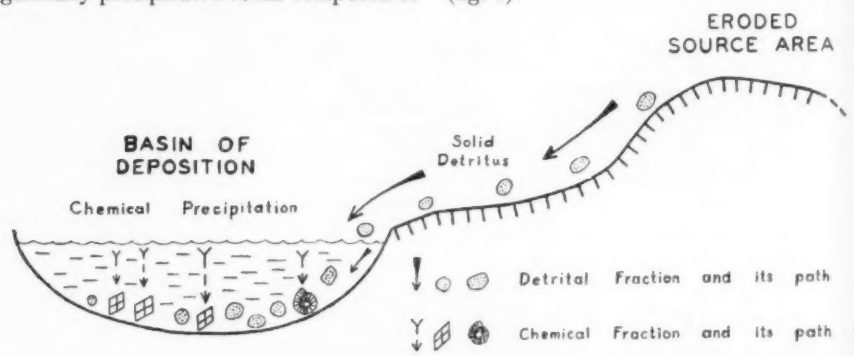

FIG. 1.—Any sediment is a mechanical mixture in every possible proportion of a detrital fraction, brought in *as a solid* into the basis of deposition from the source area, and a chemical fraction, precipitated *from solution* within the basin of deposition itself.

DETRITAL AND CHEMICAL ROCKS

All sedimentary rocks are mixtures of detrital and chemical material. For megascopic purposes detrital material is defined as consisting of *clastic* silicates. The term "silicate" is used in the extended, so-called "modern," sense and includes both minerals of the free-silica group (quartz, etc.) and the combined silicates. If the detrital material exceeds 50 per cent, the rock is named a "detrital rock"; if it is less than 50 per cent, then the rock is a chemical rock (fig. 2). If the detrital material in a chemical rock fluctuates between 5 and 50 per cent, the chances are very great that this chemical rock has a clastic (or arenitic) texture. If the de-

trital material is less than 5 per cent, the chances are fair that the rock may have a crystalline texture, although a clastic texture is also very possible.

The proportions between chemical and detrital fractions may vary considerably over very short lateral or stratigraphic intervals, as shown in figure 3, which illustrates the simultaneous operation of two basic sedimentary processes and their results. The inclined line indicates the gradual decrease with distance (between points 1 and 8) of the average grain size of the mechanically deposited detritus within a given basin of deposition. This average size drops from over 0.250 mm. (medium sand) at point 1 to less than 0.002 mm. (fine clay) at point 8. The range of deposition of these clastic textural types is shown at the bottom of the diagram.

Vertical arrows show, on the other hand, the intensity of chemical precipitation of calcium carbonate at the same given points (1–8). The arrows indicate how much of the sediment at any given point consists of carbonate. For instance, at point 1 the average grain size of the detrital fraction is 0.3 mm., and the rock as a whole contains less than 2 per cent of calcium carbonate. The rock is a medium-grained sandstone. On the other hand, at point 7 the average grain size of the detrital fraction is less than 2 microns, and the rock contains over 85 per cent of carbonates; hence the rock here is an argillaceous limestone; at point 3 it is a sandy limestone; at point 5 a siltstone; etc.

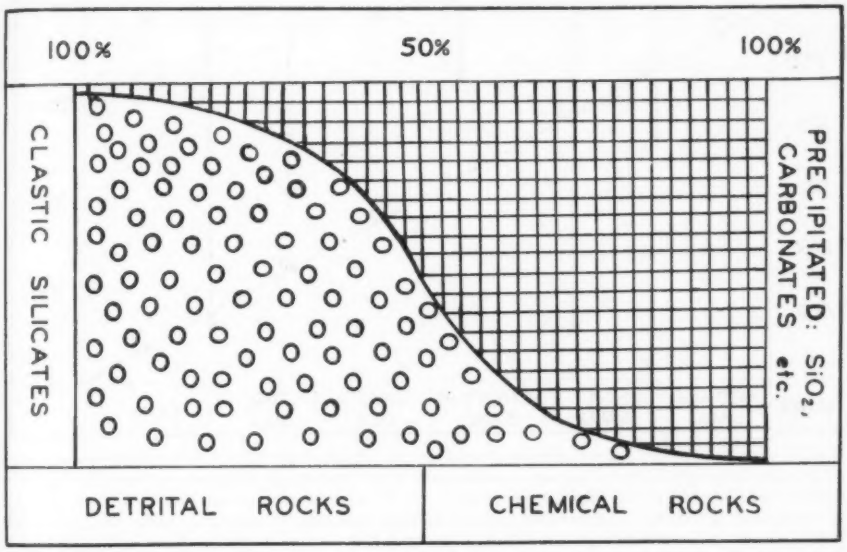

FIG. 2.—Relative amounts and character of volumetric changes between the detrital fraction (clastic silicates) and the chemical fraction (precipitated carbonates, SiO_2, etc.) in the detrital and chemical rocks, respectively.

This shows that the same two petrographic end-members, if mixed in variable proportions, may produce an almost infinite number of rapidly changing subtypes; each of these subtypes possesses

only a very local significance, and none is particularly helpful in classifying the basic sedimentary series of the region.

THREE GROUPS OF DETRITAL ROCKS

As demonstrated elsewhere (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22), the detrital fraction (fig. 4) can be divided into three groups on the basis of composition. This division has been found to hold good in rocks of all ages, from all parts of the world.

1. Quartz, with or without detrital chert grains (liberated from eroded limestones).

2. Quartz plus detrital chert grains plus abundant rock fragments (generally micaceous low-rank metamorphic rocks, such as slates, phyllites, and schists and sometimes surface igneous rocks) plus an abundance of micas and chlorite, either as large flakes or, more commonly, as a micaceous or chloritic clay or clayey paste. Feldspar may be present.

3. Quartz plus a large amount of feldspar with a subordinate (20 per cent) amount of impurities (generally similar to the rock fragments described under 2) plus or minus some clay. In this case the clay is generally kaolinitic rather than micaceous (odor test).

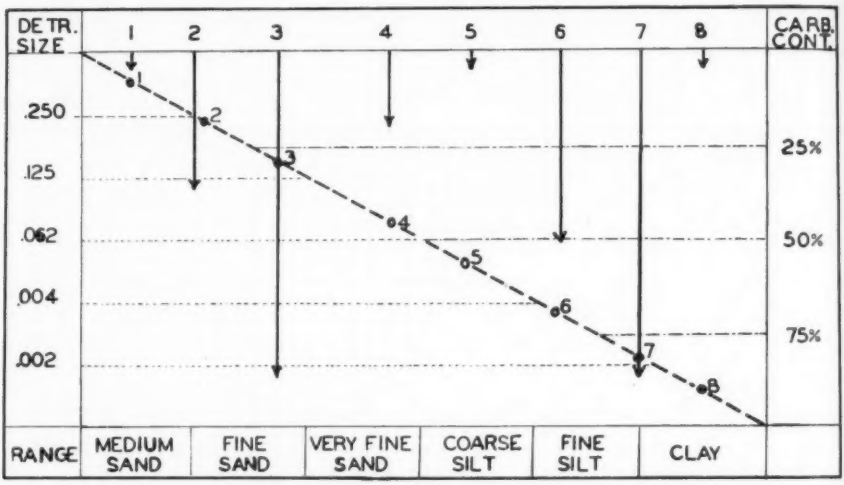

FIG. 3.—Local variations between chemical and detrital rocks over short lateral or stratigraphic intervals. *Left*, average diameter of detrital fraction in millimeters. *Right*, percentage of calcium carbonate in any given rock type (as shown by arrow). Rock types present are: (1) medium-grained sandstone (0.3 mm. av. grain size; 2 per cent of $CaCO_3$); (2) medium-grained calcareous sandstone (0.25 mm. av. grain size, 34 per cent $CaCO_3$); (3) sandy limestone (85 per cent of $CaCO_3$; 0.13 mm. av. size of detrital particles in it); (4) very fine calcareous sandstone (0.065 mm. av. size; 20 per cent of $CaCO_3$); (5) sandy siltstone (0.05 mm. av. size; 2 per cent $CaCO_3$); (6) marl (50 per cent of $CaCO_3$; 0.035 mm. av. size of detrital particles in it); (7) argillaceous limestone (85 per cent of $CaCO_3$; 0.002 mm. av. size of detrital particles in it); and (8) fine shale (0.001 mm. av. grain size; 2 per cent of $CaCO_3$).

These three mineral assemblages are used as the basic classification scheme and correspond, respectively, to the quartzite, graywacke, and arkose series of sediments as described in part 2.

THE TEXTURE OF SEDIMENTS

DEFINITION

Texture is the interrelationship of the individual particles of a rock. Texture or

"grain" can be defined as the fabric pattern of an aggregate of mineral particles.

Texture depends partly upon composition (i.e., upon types of particles and their inherent properties) and partly upon the mutual arrangement of the particles. Texture may be either clastic or crystalline.

grown within the sediment and thus have not substantially been moved after deposition and consolidation of the rock.

All detrital rocks and all impure (sandy) chemical rocks are clastic. About 80 per cent of the pure chemical rocks show a clastic texture; the balance have a crystalline texture.

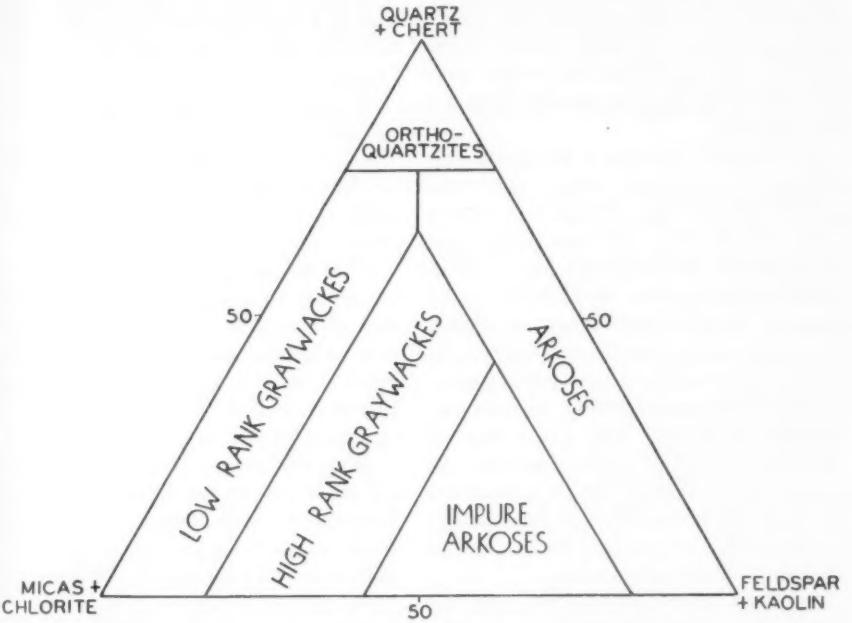

FIG. 4.—Mineral composition of the detrital fraction of the major petrographic series of sedimentary rocks. Note that the micas and chlorite may occur as rock fragments, as large, loose, individual flakes, or as a micaceous clayey paste.

1. *Clastic particles* or grains, either of detrital or of chemical origin, have been rolled around and moved to and fro before coming to rest within the sediment. They are always somewhat abraded or may be entirely broken up into fragments.

2. *Authigenic or crystalline (granular) particles* have been directly precipitated from solution on the basin bottom and have remained there or otherwise have

TEXTURAL ELEMENTS

The mineral constituents of a rock can be grouped into three main textural elements which build up the rock and give it its appearance. This grouping is on the basis of the *relative* (not the absolute) size and the relative position in space of the constituents.

1. *Grains.*—Grains or individual particles are the basic units of texture. If particles are of different sizes, occurring

in distant orders of magnitude, the term "grains" is restricted to the *larger* particles.

2. *Matrix.*—If particles are of different orders of magnitude, then the term "matrix" is used for the smaller individual units which fill the interstices between the larger grains. Hence "matrix" is a relative term only, and "matrix" by itself does not exist without the simultaneous presence of larger grains. If the particles are very small but of the same size, then they are to be termed "grains," not "matrix."

3. *Cement.*—Cement is the authigenic chemical precipitate which infiltrates around grains and matrix. In clastic rocks (which form well over 95 per cent of all sediments) grains and matrix are always clastic (either of detrital or of chemical origin), and cement is always chemically authigenic. In crystalline sediments the periods of formation of grains, matrix, and cement may overlap, although frequently the distinction of these separate generations is possible. An almost direct analogy can be worked out between the formation and morphology of crystalline sedimentary rocks and that of igneous porphyritic rocks.

All these terms are relative, and the term "grain" has no absolute size connotation.

In coarse- and medium-grained clastic rocks the possible combinations are: (1) grains (or pebbles), matrix, and cement; (2) grains and matrix; (3) grains and cement (fig. 5).

In fine-grained clastic rocks the possible combinations are: (4) grains alone (not matrix and cement or matrix alone, as frequently misunderstood by beginners); (5) grains and cement.

In crystalline rocks the possibilities are: (6) grains alone; (7) grains and cement; (8) grains and matrix; (9) grains, matrix, and cement.

CLASTIC TEXTURE AND ITS COMPONENTS

The clastic silicate fraction in a detrital rock may be loose (unconsolidated) or consolidated either by simple adhesion between the more plastic, finer-grained constituents of the matrix or by introduction of a foreign chemical cement. When this cement exceeds 50 per cent, the rock becomes a chemical rock.

Hence the two groups of detrital rocks are these: (*a*) without chemical cement (loose or consolidated only by adhesion or reorganization of the matrix) and (*b*) those consolidated through the action of chemical cements.

The two principal types of matrix are the argillaceous (kaolinitic-bauxitic) type and the micaceous-chloritic type. Both may be heavily loaded with very fine particles of quartz. A schematic representation of these two types of matrix is given in figures 13 and 14.

Both are "clayey" in the purely physical sense of the word, i.e., exceedingly fine grained. This matrix may be colored black or green by carbonaceous matter, biotite, or chlorite, or red and brown by ferric oxides. Because in this case the iron oxide acts only as a pigment and not as a cement, the term "red" rather than "ferruginous" is to be used (many black rocks contain much more iron than do some red ones).

There are two principal types of chemical cement: silica and carbonate. Their typical occurrence and relationships are shown in figure 12.

The silica cement may occur either as quartz overgrowths on quarts grains, thus greatly increasing the apparent angularity of a sediment, as chalcedony (chert), or as opal.

The carbonate cement is calcite or dolomite or, much more frequently than is generally expected, an iron carbonate of the ankerite or siderite class. Calcite is distinguished from the other carbonates by the dilute-acid test, whereas iron carbonates generally show conspicuously oxidized surfaces.

In addition, a third group of miscellaneous chemical cements includes all other chemically precipitated material. Although gypsum, anhydrite, and even halite cements may be well displayed locally, in general only glauconitic, phosphatic, and ferruginous materials are important. By "ferruginous" cement is meant clay-free hematitic or limonitic cement, which acts as an actual bonding agent and not only as a red pigment for a clayey matrix. In many cases this can be determined without too much difficulty with a hand lens.

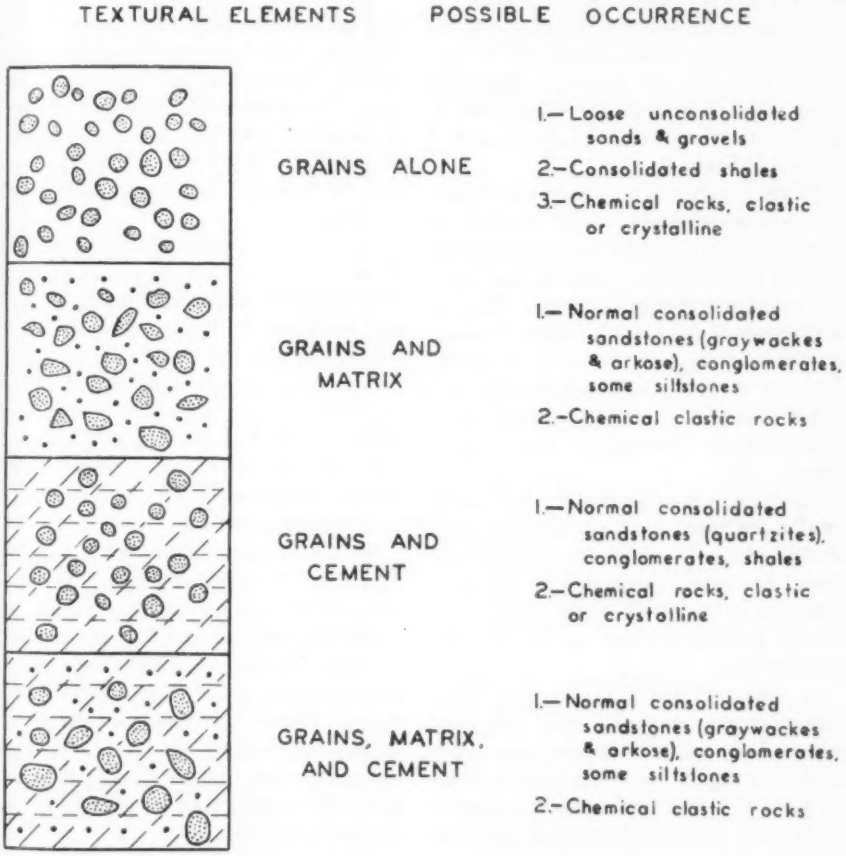
TEXTURAL ELEMENTS
POSSIBLE OCCURRENCE
GRAINS ALONE
1.—Loose unconsolidated sands & gravels
2.—Consolidated shales
3.—Chemical rocks, clastic or crystalline
GRAINS AND MATRIX
1.—Normal consolidated sandstones (graywackes & arkose), conglomerates, some siltstones
2.—Chemical clastic rocks
GRAINS AND CEMENT
1.—Normal consolidated sandstones (quartzites), conglomerates, shales
2.—Chemical rocks, clastic or crystalline
GRAINS, MATRIX, AND CEMENT
1.—Normal consolidated sandstones (graywackes & arkose), conglomerates, some siltstones
2.—Chemical clastic rocks

FIG. 5.—Textural elements and their possible combinations and occurrences

TEXTURE OF DETRITAL ROCKS

Grain size.—Detrital rocks occur in three grade sizes: (*a*) coarse grained or conglomeratic with an average size of over 2 mm., which in practice means grains of a much larger diameter in order to average up to 2 mm. with the finer matrix; (*b*) medium grained or sandy (from 2.0 to 0.0625 mm.); (*c*) fine grained or silty or clayey (average below 0.0625 mm.).

Since these textural changes occur gradually, intermediate stages can be set. These subtextures, as shown in table 3, are based on relative sizes and on the introduction into a grade size of elements from the preceding or following grade sizes. A graphic representation of the quantitative textural data of table 3 is given in figure 6.

In the conglomerate (psephite) class, common conglomerates are defined as being between 4 and 64 mm. in diameter, boulder conglomerates over 64 mm., and fine conglomerates below 4 mm. The introduction of over 20 per cent of sand, silt, or clay makes the rock a sandy, silty, or clayey conglomerate (the so-called "boulder clays," tills, and many fanglomerates belong in these categories).

Sandstones (psammites) are not easily subdivided megascopically on the basis of size, although the terms "coarse" or "fine" are easily applicable to the end-members of the sandy series (above 1 or $1\frac{1}{2}$ mm. and below $\frac{1}{2}$ mm. in diameter, respectively).

A more refined division into very coarse, coarse, medium, fine, and very fine sand is possible if the specimen under study is compared directly with a series of "yardsticks" or sand samples of known grain size, properly mounted in glass vials.

In defining grain size and especially the average grain size of a sandstone, it must be remembered that in a hand-lens or even binocular-microscope study the finer (clayey) fractions are very difficult to interpret, and hence the average grain size should be based on the size of the sandy fraction above, thus producing an average size which will always be somewhat coarser than the median diameter of a conventional mechanical analysis. These two figures should not be confused but clearly differentiated in descriptions.

In addition, depending on the type and amount of admixture, several subtextures may be established for sandstones. A rock may be termed a "normal," "conglomeratic" (over 20 per cent of pebbles), "pebbly" (over 10 per cent of pebbles), "silty" (over 20 per cent silt), or "clayey" sandstone (over 20 per cent clay).

In the finer clastics (pelites) the siltstones are described megascopically as gritty, and the shales and clays as smooth and unctuous to the touch. Sandy siltstones contain over 20 per cent of sand. Sandy shales on the basis of the preceding definition are impossible (since such "shales" would be gritty and hence automatically be called "siltstones"). Silty shales (semigritty or not perfectly unctuous), however, do exist.

Siltstones occur in two definite fabric patterns: (*a*) siltstones proper, consisting of relatively well-sorted silt particles, and (*b*) microconglomerates (or rather microbreccias), made up of relatively coarse sand grains in a very fine silty or clayey matrix. These two types, however, would be difficult to distinguish megascopically but for the fact that siltstones proper are generally somewhat more micaceous than are microbreccias.

Grain shape.—Coarse clastics can be divided on the basis of their angularity into conglomerates proper (rounded) and

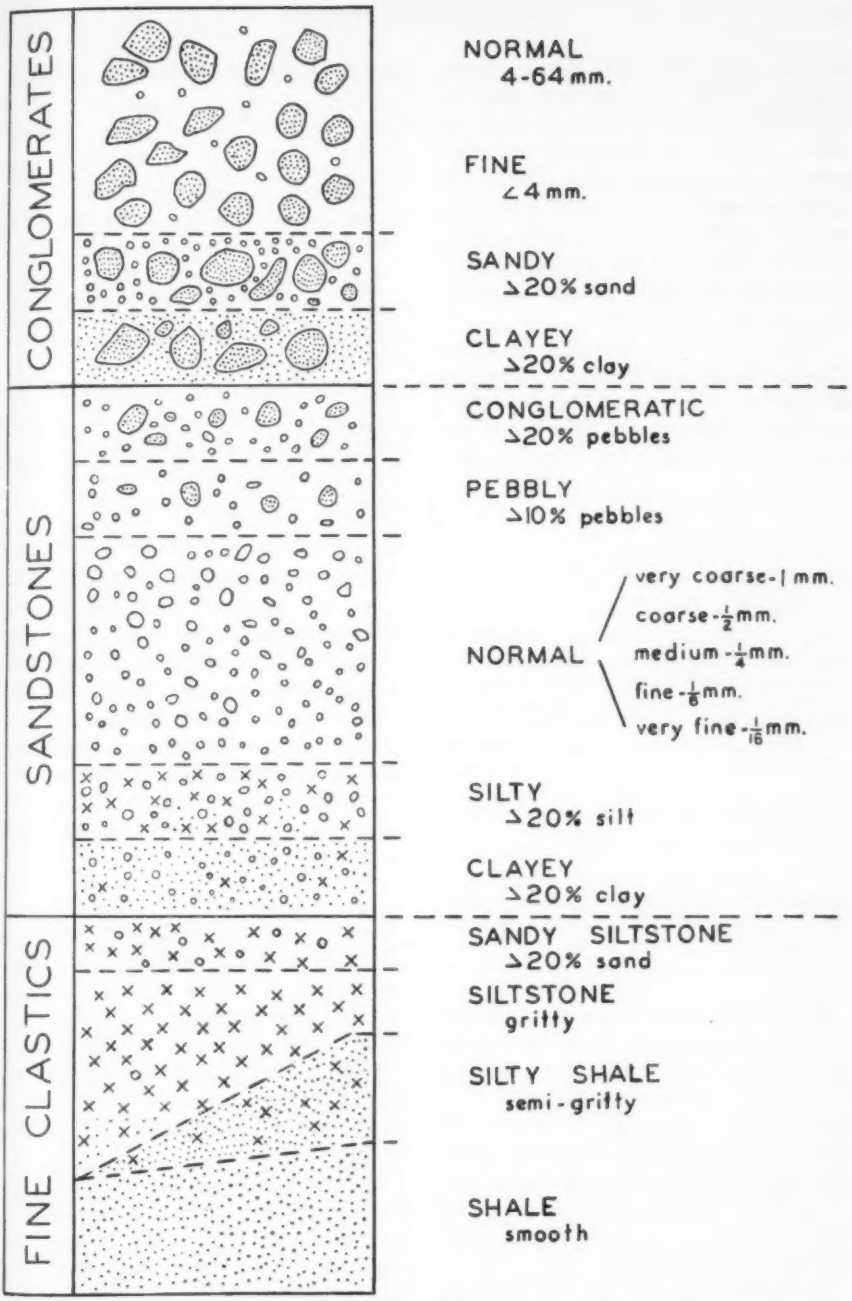

FIG. 6.—A megascopic textural division of clastic rocks, according to grade size

breccias (angular). In the medium-grained clastics this division may be somewhat difficult because many so-called "grits" are produced by overgrowths of secondary silica. Hence the terms "rounded" or "angular" should be used for sands rather than such specific names as "grit." All fine-grained clastics are angular, but this fact cannot be easily observed with a hand lens.

With a little effort it is possible to describe most sands as angular, subangular, subrounded, or rounded. This can be accomplished either through the use of appropriate comparison specimens or simply on the basis of the following arbitrary definitions, as shown in figure 7:

FIG. 7.—Definitions of roundness in clastic particles: *A*, angular: all edges are sharp; roundness coefficient is 0. *SA*, subangular: one-third of edges are smooth; roundness coefficient is 33. *SR*, subrounded: two-thirds of edges are smooth; roundness coefficient is 66. *R*, rounded: all edges are smooth; roundness coefficient is 100.

Angular—all edges of a grain are sharp.
Subangular—one-third of the edges are smooth.
Subrounded—two-thirds of the edges are smooth.
Rounded—all edges are smooth.

By assigning the values of 0, 33, 66, and 100 per cent to these respective terms and by multiplying these values by the number of grains of each class within one specimen, it is possible to describe the roundness of a sandstone in quantitative numerical terms, even on the megascopic or at least binocular-microscope level. This procedure is perfectly feasible for descriptions of well logs.

Roundness should not be confused with sphericity. At the megascopic level the sphericity of a grain can be defined as follows:

Equant—length of grain is less than $1\frac{1}{2}$ times its width.
Elongated—length $1\frac{1}{2}$ to 3 times its width (prismatic) or thickness (tabular).
Acicular or platy—length is over 3 times width or thickness, respectively.

Sphericity (i.e., the ratios between length, width, and thickness) is very difficult to evaluate correctly in particles of sand size or finer, since there may be considerable differences between the appearance of a solid particle when viewed in three or only two dimensions. Hence sphericity terms should be used with discretion at the megascopic level.

TEXTURE OF SANDY CHEMICAL ROCKS

In the chemical rocks clastic texture always occurs in the sandy or arenitic types (containing from 5 to 50 per cent of clastic silicates). The grade sizes used (coarse, medium, or fine) are the same as in the detrital rocks. If necessary, subtextural terms of the same kinds as those used for the detrital rock should be introduced for the sake of precision. The crystalline constituents should be described separately.

TEXTURE OF PURE CHEMICAL ROCKS

In the pure chemical rock (less than 5 per cent of clastic silicates) the texture may be either clastic or crystalline. If the rock contains more than 10 per cent of fragmental particles of any kind (fossil shells, intra-formational pebbles, oölites, etc.), then very probably the texture is clastic, and the same terminology is used as before.

However, the large amount of crystalline cement present should be carefully described according to the terminology suggested in the next paragraph. The rock is then referred to as a "clastic limestone," "clastic dolomite," etc.

If the texture is definitely crystalline, then the rock is no different texturally from an igneous rock. To avoid confusion, however, the terms "phaneritic" and "aphanitic" should not be used. Instead, the terms "coarsely crystalline" (over 4 mm. in diam.), "medium crystalline" (1–4 mm.), and "finely crystalline" (below 1 mm.) are suggested.

Furthermore, the degree of homogeneity of the size distribution of the crystalline elements (granules) should be expressed by the terms "even" or "uneven" (or heterogeneous), with adequate quantitative notation as to the amount of crystalline particles in each major grade size. The uneven or heterogeneous texture of crystalline sediments corresponds to the porphyritic texture of igneous rocks.

Finally, the shape of the crystalline particles should be noted, primarily their degree of sphericity (as suggested above), and their idiomorphism.

A graphic presentation showing differences between clastic and crystalline textures in chemical rocks is given in figure 15.

THE COLOR OF SEDIMENTS

The color of a sediment is the most obvious, most striking, and most easily observable property. A purely objective approach capable of yielding reproducible results is possible only on the basis of a color dictionary or at least of simplified color charts. A very fine study of color in rocks, discussing all previous work and giving a complete list of references and suggestions, has recently been made by De Ford (1944, pp. 128–137) and should be read by all geologists.

An extremely simplified summary of color in rocks follows.

1. Any color is a mixture of two basic elements: a pigment and a neutral background, which dilutes the pigment. The proportion of pigment to neutral background is known as "saturation," "purity," or "chroma."

2. A pigment is a mixture of several colors, principally blue, green, yellow, and red. The hue depends on the proportions of these four basic colors.

3. The neutral background is a mixture of white and black, in practice, therefore, being finally some shade of gray. The relative proportions of light and black will determine the lightness (or brilliance or value) of the color.

These items are usually plotted on a triaxial ellipsoid, with a vertical axis representing the white and black poles of the neutral background cutting across a central circular cross section, along the periphery of which are disposed the different hues.

This scheme can be greatly simplified by showing these relationships in two dimensions only, as sectors of a circle on a so-called "pie-shaped" diagram, as shown in figure 8.

Figure 8, *A*, shows the purity (saturation or chromatic value) of a color. In the left portion the color will be vivid or deep (high porportion of pigment); in the right portion the color will be drab or grayish (low proportion of pigment to background).

Figure 8, *B*, illustrates the difference between a light and a dark color as conditioned by the change in proportion of black to white within the neutral background.

Figure 8, *C*, shows the formation of different hues by mixing different amounts of red and yellow within the pigment.

Figure 8, *D*, reveals the combined effect of all these factors by showing the difference between orange (in which pigment predominates over background

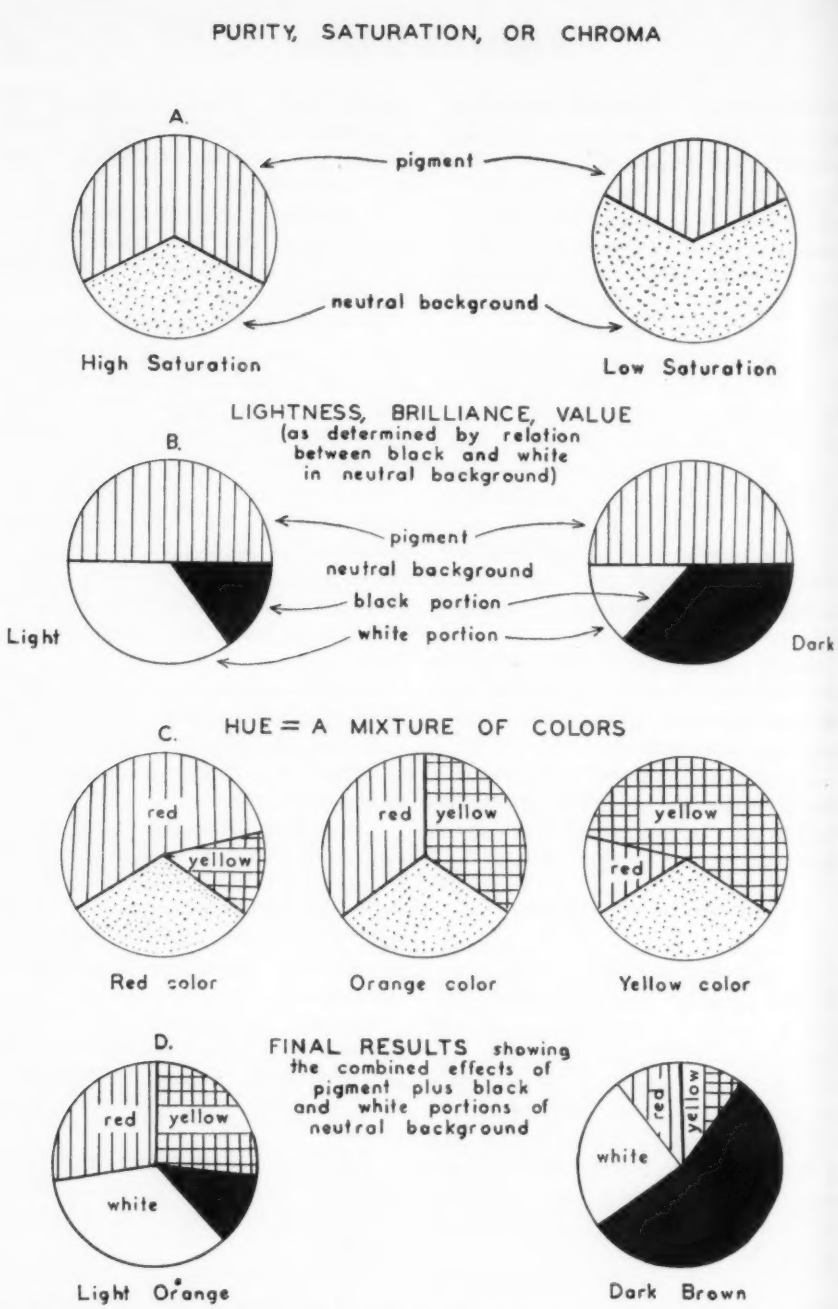

FIG. 8.—A two-dimensional interpretation of color showing: *A*, purity (ratio of pigment to background). *B*, brilliance (ratio of white to black in background). *C*, hue (ratios of basic colors in pigment). *D*, final results of interplay between *A*, *B*, and *C*.

and white over black within the background) and brown, in which the proportions of background to pigment and of black to white within the background have been reversed.

An understanding of these simple fundamentals of color is necessary to enable one to describe and compare intelligently various colors in rocks.

For rapid megascopic examination of a preliminary nature a simple approach to the description of color in rocks is possible on the basis of four basic colors and eight intermediate hues plus two neutral shades (light and dark).

The following color scheme has been found to be applicable without undue confusion or loss of time in the field:

1. The basic hue sequence consists of blue, green, yellow, and red.

2. The neutral shade sequence is light and dark, which when taken alone produced the sequence

white—gray—black.

When applied to pigments this sequence will be

light color—drab color—dark color.

3. The intermediate hues are

Between blue and green—greenish blue and bluish green
Between green and yellow—yellowish green and greenish yellow
Between yellow and red—orange
Between red and green—purple
Between red and blue—lavender and purple

These colors and hues are based not only on a study of color in the abstract but, to some extent, on the mineralogical basis of color. For instance, purple is a typical color produced by a mixture of green and red minerals (such as hematite and chlorite or hematite and glauconite) and can be used as a criterion for stratigraphic and genetic purposes.

Each of these four basic and five intermediate hues can be qualified by the terms "light" and "dark" (no qualification for the normal shade).

An additional terminology is necessary when dealing with brown and red rocks, even at the megascopic level. The normal color sequence in this case is red, orange, yellow. If this pigment is highly diluted by a neutral background, in which the black portion predominates, the following sequence of three additional colors is arrived at: cherry red, brown (or drab, depending upon relative lightness), olive.

If these fundamentals are kept in mind, then color can be used as a powerful tool in describing rocks because the variations in "shade" reflect variations in mineral composition. It must be kept in mind that a finer-grained rock will always show a more vivid color than will a coarser-grained one because of the better distribution of the pigment throughout the rock.

For more precise work the standards used in soil and rock nomenclature as reviewed and defined by De Ford should be applied.

Color should always refer to a dry fresh surface with a degree of roughness or smoothness characteristic of the rock. Wet, polished, weathered, chipped, or other non-natural surfaces give abnormal colors and should always be identified as such if used for color determination.

THE STRUCTURE OF SEDIMENTS

DEFINITION

External and internal morphology.—Structure in sediments may be defined as the visible expression of nonuniformity in texture and composition and hence is, strictly speaking, a second-order property derived from these fundamental two. However, since structure is a pri-

mary and fundamental attribute of large sedimentary bodies or rock masses, it can be considered as the *third basic property* of sediments.

Different types of structure, i.e., structural elements, or "structures" for short, can be classified as follows: (1) *morphologically* into external and internal or (2) *genetically* into primary and secondary.

The *external morphology* of a sedimentary body can be defined by its extent (size), its shape, and the character of its boundaries. Since the size and shape of a sediment are directly related to its composition and texture and since these properties can be tangibly expressed in quantitative terms, a quantitative classification of sedimentary bodies based on external morphology is also proposed. This classification is for large-scale field study and obviously does not apply to hand specimens.

SIZE OF SEDIMENTARY BODIES

The data expressing the size of a sedimentary body include the horizontal *extent* (length, width, area), *thickness*, and *volume*. Quantitative definitions of these terms are given in table 2.

In stratigraphic work very frequently these size terms (large or thick, small or thin) are used in a purely subjective way, depending upon the relationships existing within a given area or the training and idiosyncrasies of the worker. A "thick" formation within the limestone series of the Mid-Continent Paleozoic may become a very "thin" one in the arkoses of the California Tertiary. Hence objective, reproducible, and understandable terms like those of table 2 become necessary.

SHAPE OF SEDIMENTARY BODIES

The shape of a body of deposition can be expressed either in terms of configuration or through the relationships existing between the different size factors.

A. The external *configuration* of the body is described by such terms as "lens," "cone," "fan," "delta," and various types of prisms (from bars and channels to so-called "blanket" and "sheet" deposits). The meanings of these purely descriptive terms have been discussed and established adequately in textbooks on geomorphology and sedimentation, and the student is referred to these.

B. The *relation between area and volume*, another fundamental shape factor of a sedimentary body, can be expressed best as the *ratio between width* (i.e., extent measured across the strike) *and thickness*. On this basis, regardless of curvature or external configuration, there are four types of sedimentary bodies (fig. 9):

1. *Blankets*, when the ratio of width to thickness is *over* 1,000 to 1 (and up to 50,000 to 1). An example is the Lower Devonian Oriskany sand from the Appalachian region, which is over 400 miles long at the outcrop, has a width of at least 100 miles and a thickness of less than 50 feet, or a ratio of width to thickness of 10,000 to 1.

2. *Tabular bodies*, when the ratio of width to thickness is between 50 and 1,000 to 1. Examples of this are practically all the individual formations and sandy bodies in geosynclinal deposits (not their aggregates which may reach a very large thickness).

3. *Prisms*, when the ratio of width to thickness is *less* than 50 to 1 and may go as low as 5 to 1. Examples of this are most arkoses that are found in fault blocks, such as the Triassic of Connecticut (width 25 miles or less; average thickness 8,000 feet up to 16,000 feet; or a ratio of 15 to 1 or less). Prismatic deposits are typical of orogenic sediments

formed during periods of intense structural deformation of the earth's crust.

4. *Shoestrings*, when the ratio of width to thickness is less than 5 to 1, usually of the order of 1 to 1 or even less, because the thickness may exceed the width. Examples of this are most channel, many bar types of deposits, and fanglomerate belts, found within the larger tabular or prismatic bodies. The more extensive and flatter sandy bodies are generally produced by the coalescence of smaller and thicker ones. For instance, marine blanket sands represent the horizontal welding of many parallel prismatic shore lines during a long, continuous period of overgrowths (clay galls, pisolites, concretions, fossils); internal breaks (channeling, slumping features, intra-formational conglomerates, solution zones, surfaces of weathering, shrinkage, and expansion features); zones of abnormal concentration of any one of the preceding items or of such properties as sorting or porosity; structural attitude; and so on. These structural elements warrant detailed study; but, since most of these features are of no particular help in classifying or naming a sediment, they are omitted from the present discussion, and the reader is again referred to the standard text and reference books on this subject.

TABLE 2

A CLASSIFICATION OF SEDIMENTARY BODIES ACCORDING TO SIZE

Definition	Length (Along Strike) (Miles)	Width (Across Strike) (Miles)	Area (Sq. Miles)	Thickness (Feet)	Volume (Cubic Miles)
Large (or thick)	>100	>50	>10,000	>500	>500
Medium	20–100	5–50	100–10,000	100–500	1–500
Small (or thin)	< 20	< 5	<100	<100	<1

lap or off-lap. Glacial deposits of the ground moraine type and ash beds are about the only exception to this rule. However, in chemical rocks, very widespread initial bodies of the blanket type are possible.

MISCELLANEOUS STRUCTURAL PROPERTIES

Other elements of external structural morphology of sediments comprise boundaries: conformable or concordant (real or apparent), unconformable or discordant; lensing, pinching out, etc.

Elements of internal structural morphology in sediments (or "structures" for short) include layering or stratification (which should be described in terms of thickness and pattern); surface features along deposition planes (ripple marks, mud cracks, etc.); inclusions and internal

PART 2. THE THREE MAJOR SERIES OF SEDIMENTARY ROCKS

Things that have a common quality ever quickly seek their kind.—MARCUS AURELIUS.

BASIS OF CLASSIFICATION

The purpose of the present study is to introduce a purely descriptive classification of sedimentary rocks suitable for objective megascopic and field use. It so happens, however, that this classification is also genetic, since every major sedimentary series is related to a definite degree of intensity of deformation of the earth's crust (diastrophism), operating during the deposition of this rock series.

Since, however, the formulation and proof of these genetic concepts involves the presentation of a considerable body of detailed microscopic and paleogeo-

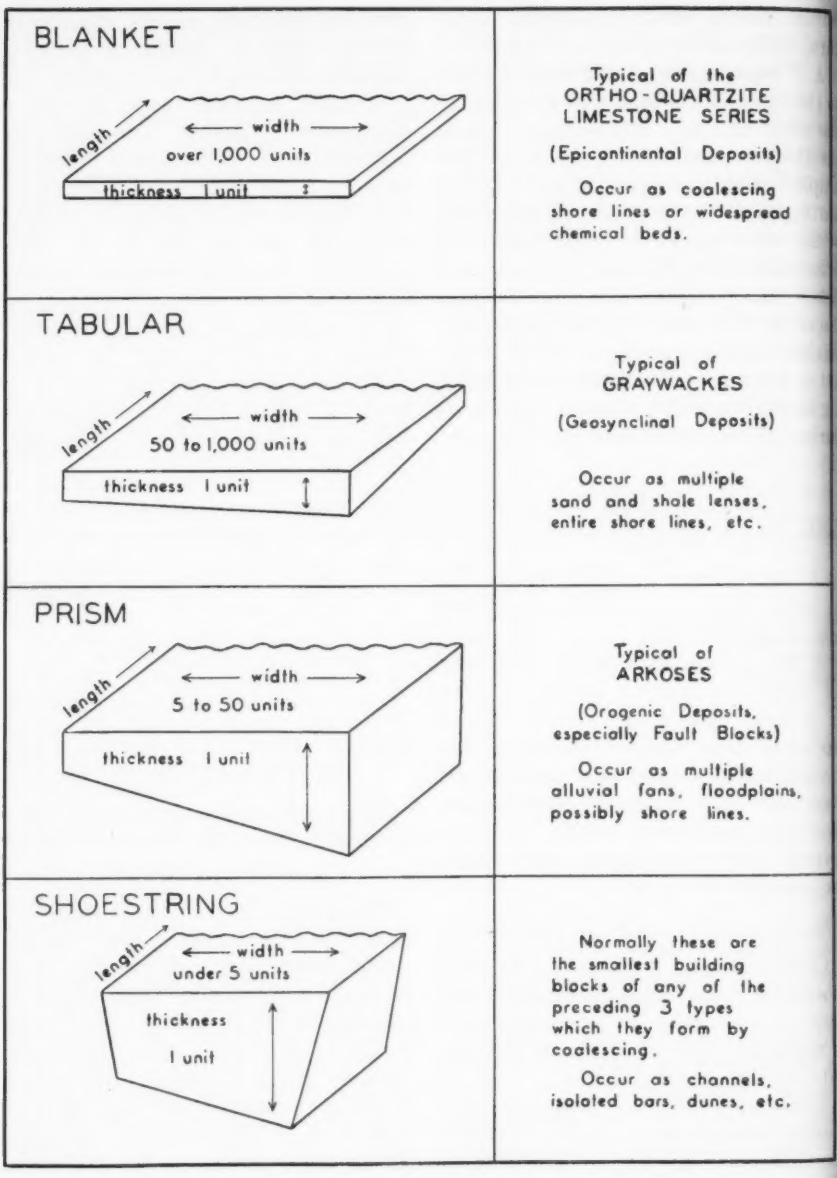

FIG. 9.—External morphology of sedimentary bodies, showing scheme of simplified geometric relationships between blanket, tabular body, prism, and shoestring.

MEGASCOPIC CLASSIFICATION

(Most of the concentrates and placers are

			TEXTURES		BONDING		C
	MAIN TEXTURE		SUB-TEXTURE		CHEMICAL CEMENT OR MATRIX		QUARTZ ±
							QUARTZITE
DETRITAL ROCKS (>50% clastic silicates)	COARSE CLASTIC (average > 2.0 mm.)	Conglomerate and Breccia class	Boulder Conglomerate (>64 mm.)		— cement (<10%)	Loose	Quartz or Ch
			CONGLOMERATE OR BRECCIA (4–64 mm.)			Consolidated A, M-C	— + QUA
			Fine Conglomerate (<4 mm.)		+ cement (>10%)	SiO_2	+ OR
			Sandy, Silty, or Clayey Conglomerate (>20% sand, silt, or clay)			CO_3	+ CHE
						Fe, G, P	— CONGLO
	MEDIUM CLASTIC (0.0625–2.0 mm.)	Sandstone class	Conglomeratic Sandstone (>20% pebbles)		— cement (<10%)	Loose	Quartz
			Pebbly Sandstone (>10% pebbles)			Consolidated A, M-C	— + "Sand rare, usually
			SANDSTONE		+ cement (>10%)	SiO_2	QUART
			Silty Sandstone (>20% silt)			CO_3	+ IMPURE (CO_3 (50–75% Si
			Clayey Sandstone (>20% clay)			Fe, G, P	— QUARTZITIC (25–50% Si + QUARTZOSE (<25% Si
	FINE CLASTIC (average < 0.0625 mm.)	Siltstone — Shale class	Sandy Siltstone (>20% sand)		— cement (<10%)	Loose	(Laterite) ←
			SILTSTONE (gritty)			Consolidated	—
			a. siltstone proper b. micro-conglomerate or micro-breccia		+ cement (>10%)	SiO_2	±
			Silty Shale (semi-gritty)			CO_3	+
			SHALE (smooth)			Fe, G, P	— +
CHEMICAL ROCKS (<50% clastic silicates)	SANDY (5–50% clastic silicates)		CLASTIC coarse: >2.0 mm. medium: 0.0625–2.0 mm. fine: <0.0625 mm.			SiO_2	Sandy and Ooli
						CO_3	+ + S
						Miscellaneous	
	PURE (not sandy) (<5% clastic silicates)		I. Clastic or arenitic if more than 10% of fragmental particles of any kind (sizes as above)			SiO_2	CHERT (nodular
			II. CRYSTALLINE coarse: >4.0 mm. medium: 1.0–4.0 mm. fine: <1.0 mm.			CO_3	+ +
						Miscellaneous	

SY

I. CHEMICAL CEMENTS

Fe — Ferruginous
G — Glauconitic
P — Phosphatic
SiO_2 — Siliceous
CO_3 — Calcareous or Dolomitic

+ and — signs indicate
end are quantitative

TABLE 3

ION OF NORMAL SEDIMENTARY ROCKS

s are omitted or left undifferentiated in this table)

COMPOSITION OF RECOGNIZABLE CLASTIC SILICATE FRACTION			
ARTZ ± CHERT	QUARTZ + ROCK FRAGMENTS + CHERT + MICACEOUS OR CHLORITIC CLAY		QUARTZ + FELDSPAR (>20%) ∓ KAOLINITIC CLAY
	− feldspar (<10%)	+ feldspar (>10%)	∓ impurities (<20%)
ARTZITE SERIES	GRAYWACKE SERIES		ARKOSE SERIES
z or Chert Gravel	Graywacke Gravel (Till— Boulder Clay)		Feldspathic Gravel (Feldspathic Till)
QUARTZ OR CHERT NGLOMERATE	Common / Graywacke Conglomerate or Breccia / Fanglomerate or Tillite	Feldspathic / Graywacke / Conglomerate or Breccia / Fanglomerate or Tillite	ARKOSIC CONGLOMERATE OR BRECCIA / Fanglomerate or Tillite
Quartz Sand	Quartz–Chert and Mica Sand ("Pepper and salt sand")		Feldspathic or Arkosic Sand
"Sandstone" , usually passes into QUARTZITE (Gannister)	gray or red Common (low rank)	black or dark gray Feldspathic (high rank)	gray or red
RE (CO_3,Fe) QUARTZITE (~75% SiO_2 in cements)	GRAYWACKE		ARKOSE (<20% impurities)
RTZITIC SANDSTONE (~50% SiO_2 in cements)			
RTZOSE SANDSTONE (~25% SiO_2 in cements)			Sub Arkose (20–40% impurities)
← Silt and Clay → (Loess) (Quartzose, micaceous—chloritic, or feldspathic—kaolinitic)			
SILTSTONE OR SHALE (Quartzose, micaceous—chloritic, or feldspathic—kaolinitic)			
Siliceous, Opaline, or Chalcedonic Shale or Siltstone			
← MARL →			
Ferruginous, Glauconitic or Phosphatic Siltstone or Shale			
and Oolitic Cherts	(rare)	Sandy Bedded Volcanic Cherts	Sandy Diatomites
SANDY LIMESTONE AND DOLOMITE (± ← oolitic → ∓)			
SANDY: Gypsum, Anhydrite, Glauconite, Phosphates, Chemical Iron Ores, Salt (rare)			
(nodular + bedded)	(rare)	CHERT (bedded)	Diatomite
LIMESTONE AND DOLOMITE (clastic or crystalline)			
SALT, GYPSUM, ANHYDRITE, PHOSPHATES, CHEMICAL IRON ORES, COAL			

SYMBOLS

gns indicate frequency of occurrence
quantitatively defined in the text

2. MATRIX

A — Argillaceous
M-C — Micaceous or chloritic

graphic evidence, it has been thought best to refrain from any fundamental discussion on this subject in the present paper. Hence the material in part 2 is designed only as a help for an easier understanding and assimilation of the classification table (table 3). The present article presents a purely descriptive megascopic classification scheme and not genetic concepts. Some of the basic ideas underlying the explanation of these genetic relationships have been discussed in a preliminary way elsewhere (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22) and are going to be presented in considerable detail in a forthcoming publication.

The present study aims at a classification of sedimentary rocks sufficiently detailed for an understanding of the relationships among sediments observable with a hand lens. This classification is essentially founded on mineral composition (reinforced by texture) and is presented in table 3. It is based, first, on the preliminary division of sediments into two main classes—detrital and chemical—and, second, on the fundamental division of the detrital fraction of either class into three main types of mineral assemblage which will correspond to the three major sedimentary series:

1. The orthoquartzite series, made up of quartz with or without detrital chert grains. Most chemical rocks also belong to this series, which can be expanded thus into a quartzite-limestone series.

2. The graywacke series, made up of quartz plus chert grains plus abundant rock fragments (generally of low-rank metamorphic rocks, such as slates, phyllites, and schists and sometimes surface igneous and volcanic rocks) plus an abundance of micas and chlorite either as large flakes or, more commonly, as a micaceous or chloritic clay. Feldspar may or may not be present.

3. The arkose series, made up of quartz plus large amounts of feldspar with a subordinate (20 per cent) amount of impurities, generally similar to the material described under 2. The clay in the arkoses is generally kaolinitic rather than micaceous or chloritic, this being readily determined by an odor test.

A diagrammatic representation of the relations existing among the major mineral constituents within the sandstone member of these main sedimentary series is given in figure 10.

Figure 11 shows the total mineral composition of four typical representatives of the sandstone members of these classes. The data of Figure 11 are given without differentiating between the detrital and the chemical components within any of the mineral groups.

QUARTZITE SERIES

The quartzite series of sediments can be popularly defined as being made up of "clean sands." It contains one detrital (clastic) end-member, consisting almost exclusively of quartz, occurring generally as medium-sized grains. These quartz grains (fig. 12) are cemented by at least one chemical end-member (cement), generally secondary silica or a carbonate (dolomite or calcite). The quartz grains are well rounded and excellently sorted.

The term "quartzite" refers to an orthoquartzite or primary sedimentary quartzite, as contrasted with "metaquartzite," which is a metamorphic rock. The vast bulk of quartzites in the earth's crust are orthoquartzites rather than metaquartzites.

Orthoquartzites, regardless of the crystallinity of their cementing material are normal sediments and have no more metamorphic significance than does a deposit of rock salt or gypsum (which are also highly crystalline rocks). The er-

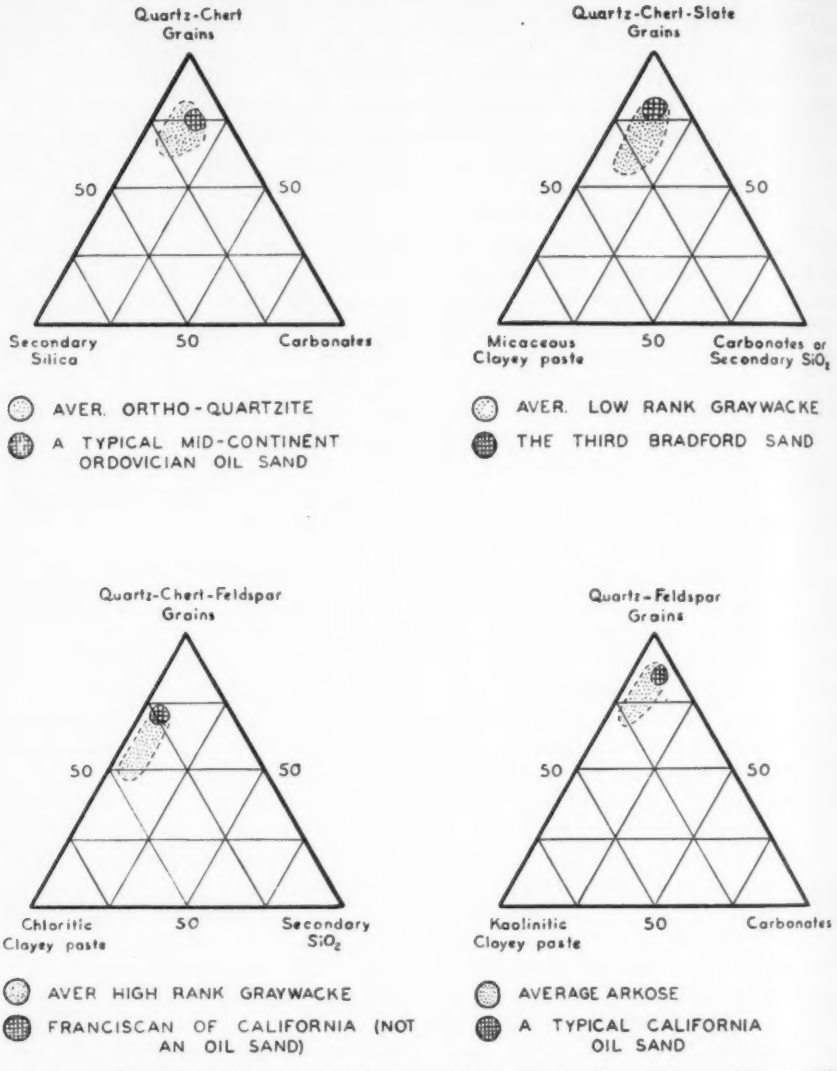

FIG. 10.—Mineral composition in terms of petrographic end-members for average sandstones and for some typical oil sands related to these sandstone classes, showing their respective fields of occurrence plotted on ternary diagrams for the orthoquartzite, graywacke, and arkosic series.

roneous and, unfortunately, fairly widely held misconception that most—if not all—quartzites are metamorphic rocks has adversely affected geologic judgment in evaluating the petroliferous possibilities of many a sedimentary province.

A so-called "textbook" quartzite in the restricted sense is a rock completely and solidly cemented by secondary quartz, so that the rock breaks across the grain (since the cementing matter is just as tough as the grains themselves).

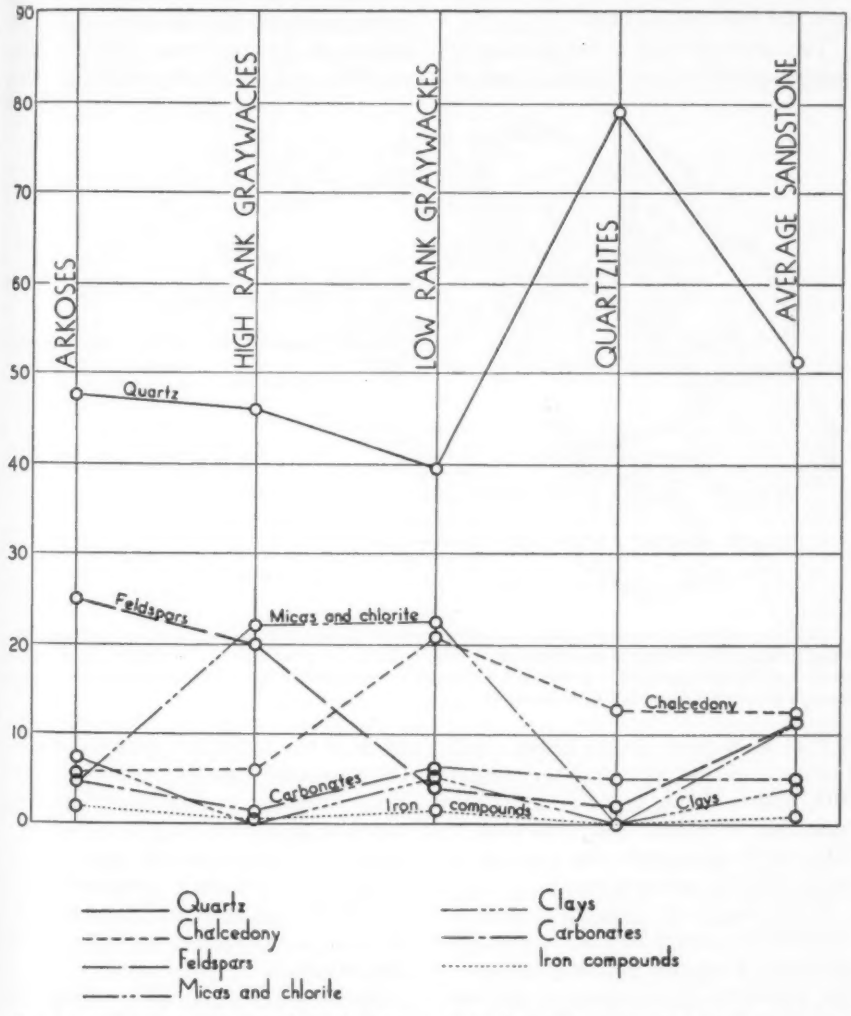

FIG. 11.—Average mineral composition of sandstones in the three major sedimentary series. Percentages shown refer to *total* composition and include both detrital and chemical constituents.

However, even the hardest and most siliceous quartzites contain some carbonate cement as an additional petrographic end-member, and, as the proportion of this nonsiliceous cementing material increases, the orthoquartzites begin to show more diversification in composition and less cohesiveness.

The fabric pattern of the average *oil-bearing* quartzite or quartzitic sandstone is that of rounded quartz grains rigidly but incompletely cemented by secondary SiO_2 (there is no porosity in a pure quartzite, 100 per cent cemented by SiO_2) with some additional parts of the pores filled by carbonates.

Structurally, the quartzitic sands occur most commonly in thin but extensive blankets. They are generally related to the quiescence (peneplanation and near-peneplanation) stage of diastrophism. The average grain size of an orthoquartzite is between 0.5 and 0.125 mm.

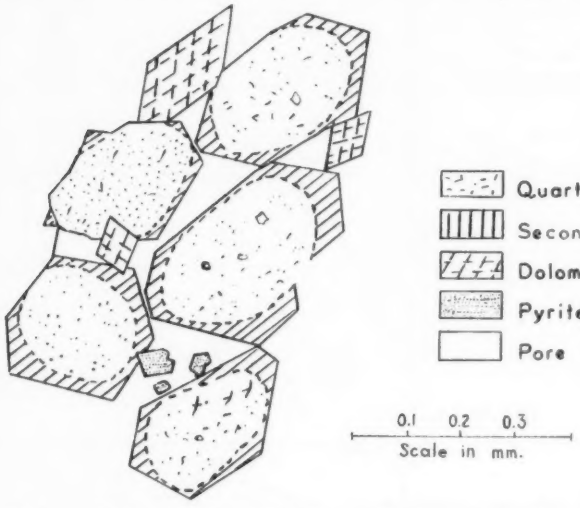

FIG. 12.—Schematic representation of an orthoquartzite (magnification 80×) showing development of pseudo-angularity through the appearance of secondary quartz overgrowths around originally rounded quartz grains and also the beginning of pitting and etching (at times mistaken for eolian frosting) on quartz grains through incipient replacement by dolomite or pyrite.

GRAYWACKE SERIES

The graywacke series of sediments can be popularly defined as a "dark dirty sand" or as a "pepper-and-salt sand" for the lighter-colored but strongly cherty varieties.

Graywackes (fig. 13) contain two or more major detrital clastic end-members: grains of quartz, chert, slate, schists, phyllites, etc., and a fine-grained matrix made of finely divided particles of micas (illite, sericite, muscovite, biotite, chlorite). Chemical end-members (cements) are subordinate and may be absent altogether. Feldspar is almost absent in the common or low-rank graywackes but may be abundant in the high-rank graywackes. In this case it is usually a sodic plagioclase.

The grains of a graywacke are medium to fine, generally very angular, and frequently elongated, and the sorting and sizing are very poor. The average grain

size of sandstones of the graywacke class fluctuates between 0.25 and 0.06 mm.

The fabric pattern of an average graywacke is a hodgepodge of angular, elongated small pebbles and rock fragments, quartz and chert grains, heavily loaded with mica flakes and bonded together by much so-called "clay," which, in reality, consists mostly of finely chopped-up or recrystallized micas or chlorite.

spathic graywackes are formed in narrow and rapidly subsiding geosynclines, frequently connected with volcanic activity (eugeosynclines). The graywackes are by far the most important kind of sandstone and are the most abundant single rock type within the sedimentary section.

Although graywackes are normally gray or dark in color (the term "graywacke" means a gray grit), nevertheless

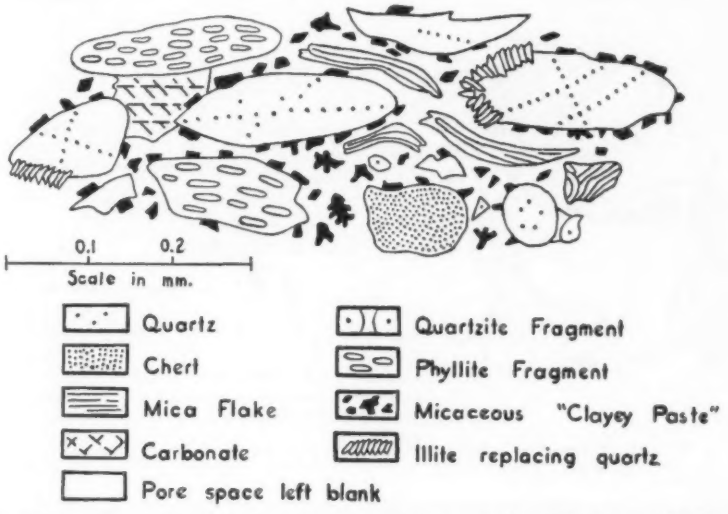

FIG. 13.—Schematic representation of a typical low-rank graywacke (magnification 125×). The diagram is an "exploded" one, i.e., the constituents have been slightly pulled apart for the sake of clarity.

Structurally, the graywackes occur as very large sedimentary bodies, thick and extensive, although the individual sandy or shaly members may be very thin and rapidly changing. Extreme lensing, channeling, and wedging-out are notable. Graywackes are formed during the stage of moderate deformation of the diastrophic cycle, the stage of subsidence, filling, and very early deformation of the geosynclines. Low-rank or common, nonfeldspathic graywackes are related to broad and gently subsiding geosynclines (miogeosynclines). High-rank or feld-

they can also be formed under the oxidizing conditions of a continental environment and in this case may be red.

ARKOSE SERIES

The arkosic series of sediments can be popularly defined as "ashy, light-gray (or red) dirty sands with much feldspathic material."

An arkose (fig. 14) usually contains one or two major detrital clastic end-members: grains derived from the rapid erosion of a granite (granitic detritus), in some cases mixed with a finer-grained

matrix of clay minerals (kaolinite-montmorillonite type) and frequently much iron oxide. Fragments of metamorphic rocks occur as subordinate impurities.

Chemical end-members are not very common but, where present, may range from carbonates (usually found) to gypsum and salt (rare).

The grains of an arkose are coarse to medium (coarser than graywackes and quartzites) and are generally angular but may often be extreme, they assume a prismatic shape.

Arkoses are related to the stage of maximum deformation of the diastrophic cycle (orogenic stage) and are formed either during or immediately after that stage. They normally follow the closing of a geosyncline and may be contemporaneous with large-scale block faulting. Many, if not most, arkoses are of continental origin and thus may be red in color. The Triassic arkose of the eastern United States can be considered the typical representative of the arkose series.

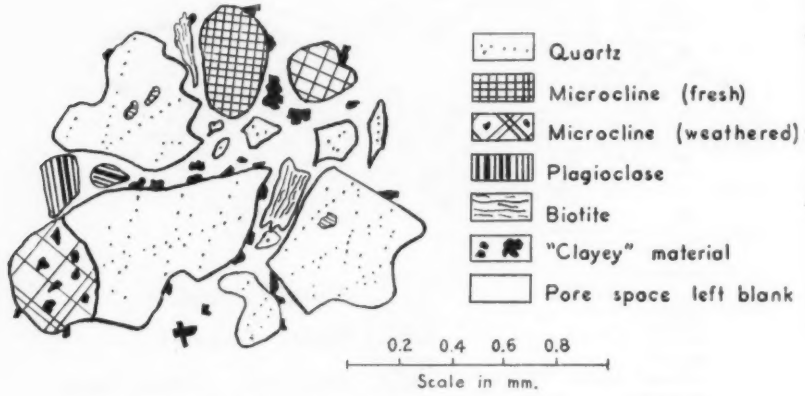

FIG. 14.—Exploded schematic representation of an arkose (magnification 40×)

somewhat more subequant in shape than they are in graywackes. The average grain size of a typical arkose is 1.0–0.18 mm.

The fabric pattern of the average arkose is a mixture of granitic detritus held together by clay, or less frequently by calcite, commonly loaded with iron oxide.

Structurally, arkoses occur as very thick sedimentary bodies, with extremely thick individual members. (The Triassic arkose is 16,000 feet thick in Connecticut.) Lensing and channeling are common, but less so than in the graywackes. Many arkosic bodies are geographically limited in extent, and thus, since their thickness

SHALES AND SILTSTONES

The fine-grained members—shales and siltstones—of each of the three major (and four minor) petrographic series discussed above are in reality mechanical mixtures made up of approximately 50 per cent of a fine silt which has essentially the same composition as the corresponding sandstone member (discussed above), roughly 35 per cent of a clay or fine mica fraction of a type which differs for each of the four petrographic subtypes, and 15 per cent of a series of

chemical cements and authigenic minerals typical of shales in general (such as phosphates and titanium oxides). These are very generalized and somewhat preliminary figures. The study of shales is difficult at the megascopic level and usually requires the use of microscopic, X-ray, and thermal analysis methods.

major petrographic types discussed above.

Most of the chemical rocks (carbonates) are related to the quartzitic series of sediments, and most chemical rocks are formed during the quiescence stage of diastrophism, which is the stage during which epicontinental deposition is at its

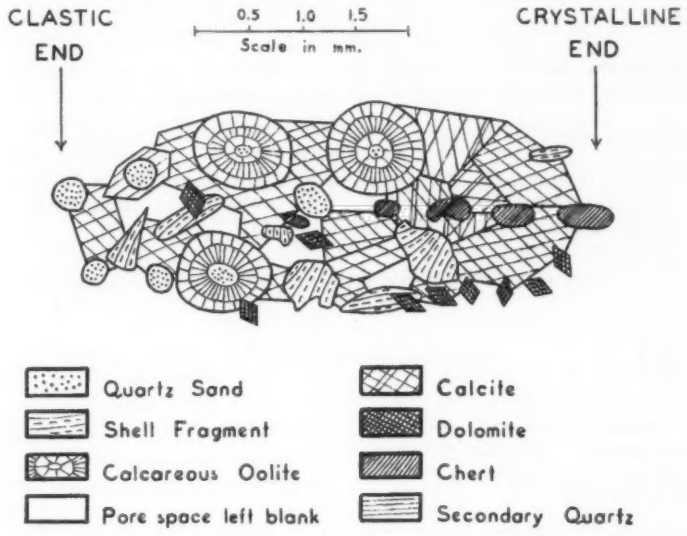

FIG. 15.—Schematic representation of an idealized chemical rock (magnification 16×) showing the passage from a sandy, clastic, oölitic limestone at the left into a densely crystalline cherty and dolomitic limestone at the right.

The four types of shales are the quartzose, micaceous, chloritic, and kaolinitic-feldspathic classes. However, it is not always possible to make such a distinction with the hand lens, although in some instances it can be done.

CHEMICAL ROCKS (QUARTZITE-LIMESTONE SERIES)

The chemical rocks (limestone, dolomites, cherts, gypsum, salt, phosphates, etc.) usually carry a certain load of detrital impurities which frequently make it possible to fit them into one of the three

peak—hence the proposed term "quartzite-limestone series," which indicates a normal and very common association between these two rock types (fig. 15).

DISTRIBUTION OF SEDIMENTARY TYPES IN THE STRATIGRAPHIC COLUMN

On the basis of a rather comprehensive petrographic and stratigraphic survey of the sediments of the United States, reinforced by the study of several fairly representative rock suites from western Europe, India, Alaska, Burma, and South America and a review of the

meager literature available on the subject, a preliminary estimate of the relative volumetric distribution of the various sedimentary series has been made.

These data are presented in table 4. The figures are believed to be of the correct order of magnitude, but no claims as to their final precision can be made at the present time.

TABLE 4

VOLUMETRIC DISTRIBUTION OF SEDIMENTARY ROCKS IN THE EARTH'S CRUST

	Quartzite Series	Graywacke Series		Arkose Series
Conglomerates and sandstones	Orthoquartzites 9 per cent	Low-rank graywackes 14 per cent	High-rank graywackes 4 per cent	Arkoses 13 per cent
Siltstones and shales	Quartzose shales 2 per cent	Micaceous shales 21 per cent	Chloritic shales 6 per cent	Kaolinitic shales 13 per cent
Chem. rocks	Limestones, dolomites, cherts, etc. 18 per cent			

PART 3. PRINCIPLES OF APPLIED PETROGRAPHIC NOMENCLATURE AND USE OF THE PROPOSED CLASSIFICATION TABLE

In practical matters the end is not mere speculative knowledge of what is to be done, but rather the doing of it.—ARISTOTLE.

PRINCIPLES OF IDENTIFICATION

The adequate description of any rock requires the naming of its mineral constituents and the definition of its texture. The degree of precision attained depends upon the character of the tools employed: megascopic or microscopic examination.

Proper identification of any rock is *impossible* without the *correct* identification of the principal mineral constituents. In such an incomplete case the classification at best can be based upon texture, bulk chemical or mineral composition, suspected mode of origin, or genetically active medium. As stated before, none of these methods is capable of yielding reproducible objective results or of conveying exactly in unambiguous petrographic terms what is meant by the name employed. Furthermore, the description and definition of texture *must* also be based on the use of objective and quantitative terms.

THE PETROGRAPHIC CLASSIFICATION OF IGNEOUS ROCKS: THE MAIN NAME AND ITS QUALIFIERS

A simple and petrographically adequate terminology is used for the classification of igneous rocks. This system is based on two major types of texture (crystalline and glassy as against clastic) and four minor types (crystalline-phaneritic, crystalline-aphanitic, glassy, and, finally, fragmental or clastic) as plotted against two major groups of diagnostic essential minerals (feldspathic versus ferromagnesian) and eight minor groups (quartz plus orthoclase, etc.).

A certain combination of texture and composition determines what is known as a "major" rock type. The name of such a rock type implies both a certain definite composition and a certain typical major texture. For instance, a granite is

a phanerite made of quartz and orthoclase.

These basic rock names are further qualified at the megascopic level by certain subtextures and by the presence of varietal minerals. The subtextures of igneous rocks are based on the degree of homogeneity (porphyritic or nonporphyritic) and on the grade size (coarse, medium, or fine) within the phaneritic textural class. Furthermore, color is given if necessary.

Hence an adequate description of an igneous rock at the megascopic level reads something like this: (1) a white, medium-grained, hornblende granite, or (2) a pink, subporphyritic, two-mica granodiorite, or (3) a greenish, very coarse, augite diorite-porphyry, or (4) a reddish quartz-felsite.

The final term or main name (granite, diorite, felsite, etc.) indicates both the dominant mineral composition and the prevailing texture. The other terms qualify the name and exactly describe the specimen. Certain main names, such as "pegmatite," may have a definite genetic connotation. A similar system can be applied to sediments.

THE PETROGRAPHIC CLASSIFICATION OF SEDIMENTS: THE MAIN NAME AND ITS QUALIFIERS

Just as in igneous rocks, it is possible to apply the following typical basic standardized descriptive sequence to sedimentary rocks: color, subtexture varietal minerals and cement, and, finally, main name.

However, sediments, as contrasted with igneous rocks, are usually characterized by a strong development of structures and in some cases may also be characteristically connected with a certain transporting medium. Hence, *if necessary*, these two additional properties may be introduced into the descriptive sequence in the following way: color—structure—subtexture—varietal minerals and cement-genetic affinity—main name. Some main names may have a definite genetic significance, such as tillite.

Examples of the classification are as follows:

1. Dark-gray, pebbly, micaceous graywacke.
2. Light-gray, thinly laminated, poorly consolidated, calcareous graywacke.
3. Red, cross-bedded, silty, siderite-bearing graywacke.
4. Red massive conglomeratic arkose.
5. White, fine-grained, rounded, dolomitic quartzitic sandstone.
6. Orange micaceous marl.
7. Pink massive feldspathic silt (or loess).
8. Gray, cross-bedded, coarse-grained, feldspar-bearing sandy dolomite.

THE MAIN NAME

The main name of a sediment should reflect both the dominant texture and the typical mineral composition. Unfortunately, since the descriptive petrographic terminology of sediments is still in its infancy, there are not enough specific terms (not even in the medium-grained detrital class) to indicate even the principal possible combinations of texture and composition.

It has been thought best to refrain, for the time being, from coining new and fancy rock names, although a tenable terminology can easily be evolved from the present classification table. This agreeable (from an author's point of view) occupation has been left to future students of the problem.

Hence some of the commoner names are used (some slightly redefined), in combination with qualifying terms based on additional information showing in greater detail the rock's composition (like the term "quartz-diorite" among

igneous rock names). This is somewhat awkward but precise.

The main name, therefore, conveys both composition and texture; for example, an arkose is a medium-grained detrital rock made of quartz and feldspar. Similarly, a sandy arkosic limestone is a chemical rock of clastic texture containing more than 50 per cent of calcite (much of it in rolled grains), with a sizable amount of actual quartz and feldspar grains; or a quartzose marl is a fine-grained detrital rock made of very fine-grained, megascopically unrecognizable, aluminum silicates, with a large amount of calcareous cement and carrying recognizable quartz grains. Finally, a quartz conglomerate is a coarse-grained detrital rock consisting of rounded quartz pebbles, whereas an arkosic breccia is a coarse-grained rock, consisting of angular quartz and feldspar pebbles. If this breccia is demonstrably of alluvial-fan origin, the term "arkosic fanglomerate" is applicable; if it is certainly glacial in origin, the term, "arkosic tillite" may be used.

The main name is modified by color, subtexture, character of bonding material, presence of important varietal minerals, and, if necessary, structure and genetic agent.

QUALIFICATION BY COLOR

The best and safest procedure in describing the color of a sediment is to use a recognized color chart. Since in practice this is generally very difficult, it is best to limit one's self to fairly objective and commonly understood names, such as those found in the basic colors of the spectrum (violet, blue, green, yellow, red), reinforced by the following additional terms: brown, olive, orange, pink, purple, white, gray, and black. Furthermore, it is permissible to combine any two colors (such as bluish-green or greenish-blue) with the second term being the dominant one. Finally, it is also possible to preface each term by the words "light" or "dark" to indicate the intensity of the neutral background (frequently kaolin in the one case, organic matter in the other). Subjective terms, such as "chocolate brown," should be avoided.

QUALIFICIATION BY SUBTEXTURE: GRADE SIZE AND ANGULARITY

IN DETRITAL ROCKS

The following scheme is proposed for detrital sediments:

1. In the coarse-grained rocks the main name is the *XYZ* conglomerate, breccia, tillite, or fanglomerate. The *XYZ* refers to composition. This main name should be prefaced by the term "boulder," "fine," "sandy," "silty," or "clayey" if necessary. When normal and uncomplicated (i.e., 4–64 mm. in diam. with no finer admixtures), the rock is just a normal unqualified conglomerate, breccia, tillite, etc., of a definite composition.

2. In the medium-grained classes the main names, such as quartzite (or quartzitic or quartzose sandstone), graywacke, or arkose should be prefaced by the qualifiers "conglomeratic," "pebbly," "silty," or "clayey" if necessary. Additional terms implying a definite grain size, from very coarse to very fine (if correctly determined) should be added here.

3. In the finer-grained detrital rocks the main names "siltstone" (properly determined as to mineral composition) or "loess" are to be prefaced when necessary by the qualifier "sandy," whereas the main names—shale, clay, or marl—may be prefaced again, when necessary, by the term "silty."

The angularity or, conversely, the rounding of the constituents is strikingly seen in the coarser clastics and is such an obvious property that it should be incorporated in the name. A breccia or fanglomerate or tillite (as contrasted with a conglomerate) should contain no less than 25 per cent of definitely angular fragments (i.e., fragments with sharp edges).

In the medium-grained clastics angularity is a less noticeable property. As a rule, the constituents of all graywackes and almost all arkoses are relatively angular, whereas those of most quartzites are rounded. Pseudo-angularity produced by secondary silica overgrowths on the grains of quartzitic rocks should not be mistaken for original angularity.

The term "grit" is considered superfluous. It usually refers to a pebbly graywacke or a pebbly arkose, rocks in which the grains are normally angular, but where the increased grain size makes this angularity more obvious to the casual observer.

Where observable, the rounding or angularity should be noted by using approximate qualifiers before the main name.

IN CHEMICAL ROCKS

Two main textures are possible in the chemical rocks: clastic or crystalline.

1. In the sandy chemical rocks (i.e., those containing over 50 per cent of clastic silicates) the main name, such as sandy limestone, sandy dolomite, oölitic chert, etc., should be prefaced by the terms "coarse," "medium," or "fine" to indicate grain size.

2. In the pure chemical rocks (i.e., those with less than 5 per cent of clastic silicates) the texture is also probably clastic if over 10 per cent of the rock consists of obvious fragmental material of any kind, such as fossil shells. In such a case the same terminology as above may be applied. The main name in this case will be clastic (or arenitic) limestone and so on, and these main names will be modified by the following subtextures: coarse, medium, or fine.

3. If the texture is crystalline, the main name is limestone, dolomite, gypsum, etc., and the qualifier is "coarsely grained," "medium-grained," or "fine-grained." If the crystallinity is to be emphasized or is conspicuous (as is usually the case with coarsely crystalline rocks, which, again, are easier to describe than finer-grained ones), then the term "coarsely crystalline" may be used.

QUALIFICATIONS BY SUBTEXTURE: BONDING AGENTS

1. If the detrital rocks are loose (unconsolidated) and lack bonding, the terms "gravel," "boulder bed," "sand," "silt," or "clay" should be used.

2. If the detrital rocks are consolidated by a nonchemical matrix (either produced by simple adhesion of the original finer-grained detrital constituents or by their later reorganization), then the main name is used without additional qualification if the matrix is of the normal mineral composition for that particular rock. For instance, the normal matrix for a graywacke is micaceous or chloritic, whereas the normal arkose possesses an argillaceous, i.e., a kaolinitic matrix, and the name "arkose" presupposes such a matrix. If, on the other hand, the matrix of an arkose is abnormally, but obviously, high in fine micas or in chlorite, then the qualifier "micaceous" or "chloritic" is added to arkose.

A quartzite or a quartzitic sandstone normally has no matrix whatsoever. It is bonded exclusively by chemical cements.

Hence any matrix in a quartzitic sediment is abnormal, and in such a case the

main name, "quartzite," is qualified by the term "chloritic" or "argillaceous." The chances are that such rocks will prove to be very close to graywackes and may turn out to be the so-called "winnowed" graywackes with an abnormally high quartz content (a typical and excellent type of oil reservoir). These borderline and transitional types are going to be discussed in some detail in a forthcoming publication dealing with the genetic aspects of the problem.

If the argillaceous, micaceous, or chloritic material is red in color, the term "ferruginous" *should not be used*, since the bonding effect is produced not by the iron oxide but by the clayey material. Since in this case the ferric oxide acts only as a pigment, the term "red" should be used instead.

3. If the detrital rock is consolidated by a chemical cement, then the main name should be prefaced by the terms "siliceous," "calcareous," "dolomitic," "ferruginous," "glauconitic," or "phosphatic." In less frequent cases the main name may be followed by such expressions as "cemented by gypsum" or "by halite."

The term "ferruginous" is definitely restricted to a chemical cement consisting of ferric oxide. Even then the term is not absolutely satisfactory, and it would be much better to employ the term "hematitic" or "limonitic" instead; but this may be asking too much of the average field geologist without a streak plate. It should be pointed out, nevertheless, that the term "ferruginous" is a gross misnomer when used promiscuously to designate red rocks in general. Indeed, many dark rocks, containing iron in the ferrous state, may have more iron than do some red rocks in which the iron is in the ferric state. In igneous rocks the term "ferruginous granite" is immediately recognizable as a patent absurdity, whereas the term "biotite or hornblende granite" indicates that the iron is there and also how it occurs. The same elementary care in definition should be employed in the description of sediments.

FREQUENCY OF DISTRIBUTION OF SEDIMENTS IN TERMS OF BONDING MATTER VERSUS DETRITAL MATTER

FREQUENCY OF BONDING AND CEMENTING MATERIAL IN DETRITAL ROCKS

In theory any detrital rock can occur with any kind of cement; in practice this is not so. It has been found empirically that certain mineral assemblages in the detrital fractions have a definite tendency to occur with certain types of chemical cements or may lack chemical cements altogether. The genetic reasons for these variations are given in some of the references already mentioned (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22).

The sympathetic or antagonistic associations of certain detrital fractions with certain types of chemical cement are shown by the plus (+) signs when common and the minus (−) signs when rare. These signs indicate the degree of probability for the existence of a given rock with a given type of bonding matter and chemical cement. For instance, the minus (−) sign opposite the CO_3 symbol in the feldspathic graywacke column shows that this rock is very rare. Indeed, calcareous high-rank graywackes are almost nonexistent. On the other hand, the plus-minus (±) signs opposite the CO_3 symbol in the common graywacke column shows that the probability of its occurrence is better than fair; and, indeed, calcareous low-rank graywackes are frequently found (for instance, the so-called "molasses"—in the petrographic sense—of the European geologists are just that).

The symbols have the following approximate values or order of magnitude:

‡ = Always present, about 90 chances out of 100 for its occurrence.
+ = Normal type of occurrence, excellent probability of being present, from 70 to 90 chances out of 100.
± = Fair to good chances of being present, about 50–70 out of 100.
∓ = Rather poor chances of occurrence, about 20 to 50 out of 100.
− = Very poor chance of being present, less than 20 out of 100.

These percentages refer to frequency of occurrence within the same rock type. For instance, in the case of common low-rank graywackes the plus (+) sign in the *A, M–C* tier indicates that of 100 low-rank graywacke specimens perhaps 70–90 per cent will carry as the predominant bonding medium (either alone or in combination with subordinate cements) an argillaceous or rather micaceous chloritic matrix. Furthermore, the plus-minus (±) sign in the carbonate tier of the same graywacke class indicates that of the same 100 low-rank graywacke specimens 50–70 may contain sizable amounts of carbonate cement, and so on for the other bonding media.

The use of plus and minus signs is for the guidance of the student and shows him what the odds are for the possibility of the occurrence of a given rock type and thus serve as a rough check on his identification of the rock.

The use of the plus and minus signs in the finer-grained detrital rocks is similar to that in the coarse and medium-grained class. For instance, the minus sign opposite the SiO_2 symbol in the siliceous or opaline shale tier in the low-rank graywacke column indicates that such siliceous shales are rather rare where they carry a coarsely detrital fraction of micaceous composition with no feldspar but are common (+) if feldspar appears in addition to mica and chlorite (feldspathic graywackes column).

These values are admittedly very general approximations only, but their order of magnitude is, on the whole, believed to be correct.

FREQUENCY OF SUBORDINATE DETRITAL MATERIAL IN CHEMICAL ROCKS

Since the chemical fraction is dominant in the chemical rocks, the use of the plus and minus signs in this category is reversed, and they are employed to indicate the probability of occurrence of a certain type of detrital fraction within a given background of chemical precipitates.

For instance, the double plus (‡) signs in the quartzite column of the sandy limestone class indicates that sandy limestones generally contain quartz (or chert) grains and, conversely, that they are almost nonexistent with a detrital load corresponding to that of a high-rank graywacke (−) but that they have a better possibility of carrying an arkosic load (∓) and a very good possibility of carrying a low-rank graywacke type of detritus (+).

The plus and minus signs left and right of the oölitic subcaptions show that the probability of oölitic occurrence increases as the silicate detritus assumes a more purely quartzose character.

QUALIFICATIONS BY VARIETAL AND ACCESSORY MINERALS

The essential minerals which determine the main name of a sediment are specifically listed under the heading of "Composition." Within the detrital rocks these essential minerals are quartz-chert for the quartzitic series; quartz, chert, mica, micaceous clays, and rock fragments for the graywacke series in general and also feldspar for the high-

rank graywacke subseries; and quartz, feldspar, and nonmicaceous normal clay minerals (i.e., kaolinite and bauxite) for the arkosic series of detrital rocks.

Any other mineral not specifically mentioned in this list becomes a varietal mineral if it exceeds 1 per cent in amout and hence becomes a major constituent that is readily recognizable with a hand lens. The terminology employed to indicate the presence of such varietal minerals should, theoretically at least, be different from that used to designate the occurrence of chemical cements.

A system employing a terminology such as "gypsum-bearing" or "mica-bearing" or "garnet-bearing" would have the advantage of being entirely clear but would be awkward and cacophonous and hence undesirable. As a compromise between scientific and linguistic clarity the following system is tentatively proposed:

a) Adjectives ending in *-ic* or *-ous* should be used for varietal minerals when no confusion is possible with chemical cements. For instance, in "micaceous arkose" it is clear that mica is a varietal detrital mineral (possibly in the matrix) rather than a chemical cement.

b) If confusion is possible, then the less elegant but more precise term "mineral-bearing" is to be employed. An example would be "glauconite-bearing dolomitic graywacke."

Under "accessory minerals" are included mostly the so-called "heavy" minerals. Generally these minerals cannot be studied at the megascopic level. Any mineral which occurs in amounts of less than 1 per cent in a sediment is considered to be an accessory and is not mentioned in the megascopic definition of the rock. Its presence, if determinable megascopically, is mentioned in an additional sentence when proceeding with the detailed description of the rock.

QUALIFICATIONS BY STRUCTURE

Whereas in igneous rocks a massive structure is the rule and layered or banded structures are the exception, the opposite is true in sediments. Two possibilities may exist: (1) the structure of a sediment is simple and can be defined in one or two words (cross-bedded, thinly laminated, varved, massive); or (2) the structure is complex and needs several words or an entire sentence for its definition.

In the first case the definition can be incorporated in the description preceding the main name, and its place will be between color and subtexture (for instance, a red, cross-bedded, pebbly arkose).

In the second case the definition should follow the main name and should be prefaced by the word "showing." For instance, a "dark-greenish-blue, silty, high-rank graywacke, showing well-developed cyclic graded bedding in 4-cm. bands with incipient and semimicroscopic cross-bedding within the coarser portion of each 4-cm. layer." If the description of the structure begins to be loo long, it can be separated into a sentence of its own following the basic definition of the rock.

The presence of fossils and recognizable organic remains is considered to be a structure and should be recorded at this state (example: white, coarse, cross-bedded, fossiliferous, dolomite-bearing quartzite).

GENETIC AND ENVIROMENTAL TERMS

Many rock types are produced by certain specialized sets of processes (frequently and somewhat loosely referred to as "environments"). This combination of processes may produce a certain typical combination of composition and texture and possibly structure in the resulting rock. The nomenclature of these specialized sets of processes ranges from

such generalized terms as "marine" or "continental" to much more definite terms, such as "eolian," "glacial," or "of alluvial-fan origin," etc. In the last three instances the products of these processes have been frequently called "loess" (a somewhat loose usage), "till," and "fanglomerate." Theoretically and in the abstract, any kind of mineral composition should be possible for a till or a fanglomerate. Actually this is not so because genetic processes, formative conditions, and so-called "environments" do not exist in a geologic vacuum but operate against a certain dominant diastrophic background which not only determines the probable relative intensity and effectiveness of any one process but also the type of material that it will have to work with.[1]

For instance, the rigorous climatic or topographic conditions which produce a fanglomerate, a tillite, or a loess (deflated material of a glacier or of a dried-up fluvial basin) are all very typical of the large-scale emergence of continents and of periods of relatively intense orogeny and hence are characterized by arkosic sediments, possibly diluted by some graywacke-like material. The validity of this conclusion has been pragmatically verified by a petrographic study of numerous assorted loesses, tillites, and fanglomerates. For these reasons loess and fanglomerates are tentatively defined not only as eolian or alluvial-fan type sediments but also as highly feldspathic ones.

Tillites are also normally defined as basically arkosic and hence in special cases should be qualified as graywacke-tillites. Hence these genetic terms are elevated to the rank of main names, characterized by composition, texture, and also in this case formative agents. An igneous analogy is the term "pegmatite," which also has a definite genetic meaning.

Other genetic terms, such as "eolian," "flood-plain" (deposit), "glacial," "continental," "marine-beach" (deposit), "desert" or "desertic" (deposit), "paludal," "lacustrine," etc., if considered necessary and if conclusively known to be *correct* for the particular sediments under study, can be used as qualifiers to redefine the main name. These terms should be placed before the main name if they are brief or after the main name if they are lengthy. In the latter case the main name should be followed by a statement that the specimen is a clayey graywacke-breccia (or boulder clay) probably of late glacial origin.

Such dubious terms as "silicified" or "recrystallized" (both great favorites in stratigraphy) and similar pseudo-genetic terms which imply a knowledge that the observer at the megascopic level does not (and cannot) possess, should be avoided. Use, instead, the terms "siliceous" or "crystalline," which are correct, truthful, and adequate.

USE OF PROPOSED TERMINOLOGY AND CLASSIFICATION TABLE

In summary, the following step-by-step procedure is suggested for the adequate megascopic description and classification of a sedimentary rock:

1. The rock is classified as a detrital or chemical rock, depending on the relative proportion of detrital material (in practice almost entirely restricted to *clastic* silicates) and chemical material.

2. The rock is classified further as belonging to the quartzite, graywacke, or arkosic series, depending upon the com-

[1] The principles behind this concept are explained in a general way in the references cited earlier (Krynine, 1942, pp. 537–561; 1943; 1945, pp. 12–22), and are to be described in specific detail in a forthcoming publication.

position of the clastic silicate fraction. This division when dealing with medium-grained clastic rocks (detrital rocks or sandy chemical rocks) can practically *always be accomplished successfully by a moderately well-trained observer* and has been successfully carried out without undue difficulty by upperclassmen and even Sophomores working under the writer's guidance at the Pennsylvania State College.

3. Such a division may not be possible when dealing with fine-grained clastic rocks (detrital rocks or sandy chemical rocks) or with pure chemical rocks with no recognizable silicate fraction. Then the terms "limestone," "siltstone," and "shale" are to be used in the same way as the term "felsite" is used in igneous rocks.

4. If the rock is a detrital one, after placing it in the proper series, determine whether it is bonded by a fine-grained matrix or a chemical cement. This can be accomplished, again without undue difficulty, by observing the clastic and grained character of a matrix as compared with the densely crystalline character of a chemical cement. At this stage the plus and minus signs will serve as a guide to the probability of the identification's being a correct one.

5. If the rock is chemical, note whether it is sandy (over 5 per cent of clastic silicates) or pure. If it is pure, determine whether the texture is arenitic (clastic) or crystalline.

6. If the chemical rock is sandy, determine the detrital series to which the sandy fraction belongs. This can be done with little effort in many, or most, sandy chemical rocks if they are medium or coarser grained. If the chemical rock is pure, this may prove to be impossible; but the attempt should always be made. Again the plus and minus signs will serve as a guide. If necessary, dissolve a small amount of the rock in acid and examine the insoluble residue with a hand lens.

7. Proceed with the complete identification and description of the rock as suggested in the preceding paragraphs by following the scheme of: color—structure—subtexture—varietal minerals—cement and bonding (or, conversely, detrital material in a chemical rock)—genetic affinity—main name.

CONCLUSIONS

SCIENTIFIC SIGNIFICANCE

In the present age of specialization there has been a tendency among field geologists not to bother unduly with rock specimens and to leave their detailed description to petrographers. As a result, a comparison of many recently published stratigraphic and areal geologic studies shows that, in so far as descriptions of rocks (which, after all, are the primary building blocks of all geological work) are concerned, these recent publications are less detailed—and frequently less correct—than similar descriptions published over a hundred years ago. For instance, the rock descriptions found in Percival's classic monograph on the geology of Connecticut, published in 1842, will compare favorably, both in precision and in accuracy, with most modern nonspecialized megascopic treatments of sediments.

Since all rocks are observed in the field and most of them never get past this stage of hand-lens description, it follows that a considerable improvement in megascopic standards would be most useful. In addition to the satisfaction of a duty well done, such improvement would provide, among other things, adequate material (so-called "sedimentary evidence") for a much better interpretation of paleogeographical and structural

events, an interpretation which to a large extent should be carried out in the field.

PRACTICAL APPLICATIONS

At the present time approximately 50–60 per cent of graduates in geology go into the petroleum industry. It is not an exaggeration to say that during the first few years of their careers, at least half of their time will be spent "sitting on wells," i.e., examining drill cuttings and drawing the necessary geological conclusions, which—if wrong—may be extremely expensive. On that basis it follows that the odds are strong that at least one-third of the total work done by the average geologist for five years or so following graduation will be spent on the equivalent of an applied course in elementary hand-specimen petrology, and specifically on the sedimentary portion of it. Hence, to put it mildly, the standards of such work—again in respect to both precision and accuracy—ought to be greatly improved.

Search for the so-called "stratigraphic traps" is mostly a petrographic undertaking, which, if accomplished in the field or in the average oil-company laboratory, again resolves itself into the interpretation of megascopic data. Stratigraphic traps are essentially a function of the relations existing between grains, on the one hand, and either matrix (the common detrital sand-shale lensing) or cement (equally common changes in chemical constituents), on the other. The characters of these changes are quite different in different rock series, and hence an adequate working knowledge of simple sedimentary elements and the ability to recognize them rapidly in the field are desirable. Finally, studies by the writer have shown that the three major sedimentary series are characterized by three entirely different types of oil fields and that the methods of prospecting—and exploitation—which are successful in one series may fail completely in the other.[2]

This should come as no surprise to the igneous petrographer and economic geologist, who know that tin deposits are related to granites and nickel deposits to gabbros and dunites. Similar genetic relationships exist in the distribution of different types of petroleum deposits. Such genetic relationships can be understood best when the description and classification of sedimentary rocks reaches a degree of precision at least equal to that proposed in the present article.

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[2] The seismograph, for instance, works very well in the orthoquartzite-limestone series but frequently does not work at all in the graywackes.

REVIEWS

Second Symposium on the Age of the Saline Series in the Salt Range of the Punjab Held at Udaipur on 27 and 28 December, 1945, under the Joint Auspices of the National Academy and the Indian Academy of Sciences. (*Nat. Acad. Sci. India Proc.*, sec. B, vol. 16, [April, 1947].) Pp. 1+257+ many illus. Rs. 15 to nonmembers abroad.

The so-called "Saline series," or salt marl, occurs at the base of the sedimentary sequence of the Punjab Salt Range and has long been the subject of one of the most interesting controversies in Indian geology. At most places these beds underlie the unfossiliferous Purple sandstone (believed to be Cambrian because it underlies fossiliferous Cambrian strata) or the Talchir boulder beds (late Paleozoic tillite). Stratigraphic position suggests that the Saline series is of early Cambrian or pre-Cambrian age. Nowhere has it been observed in unquestioned, undisturbed contact with older rocks.

Generally similar salt-bearing beds occur in the Kohat region west of the Indus River and are recognized by all to be of late Eocene age. Outcrops of these deposits are separated by only 17 miles from the salt marl of the Salt Range. This fact, the generally "young" appearance of the Saline series, and the presence of oil shale and shows of oil within it have naturally led to the conclusion that the salt-bearing beds of Kohat and the Punjab are of equivalent age and that the latter owes its apparent stratigraphic position to the overthrusting of more ancient strata.

These divergent views regarding the age of the Saline series and the fundamental structure of the Salt Range have been debated for many years. More recent investigations have furnished evidence interpreted as being favorable to both contentions. Many fragments of angiosperm and gymnosperm plants and insects of undoubted post-Paleozoic age have been recovered from the rock salt, dolomite, and oil shale of the Saline series. On the other hand, field studies have failed to establish a widespread plane of thrusting above these strata, and at many places the stratigraphic sequence appears to be normal and undisturbed.

The present symposium is the second to consider this problem. It followed a field excursion to the Salt Range, where a number of critical exposures were carefully examined and sampled.

The principal proponents in the controversy at the present time are Professor B. Sahni, of the University of Lucknow, and Mr. E. R. Gee, of the Geological Survey of India, who favor late and early ages of the salt marl, respectively. Professor Sahni discovered fossil fragments in the rock salt and has been instrumental in interesting several other investigators in their study. Mr. Gee has engaged in extensive field work in the Salt Range and has mapped several parts of it in detail. He formerly indorsed the view that the Saline series is Eocene, but field evidence forced him to change his mind. The positions taken by these gentlemen in general reflect the opinions of paleontologists, on the one hand, and stratigraphic and structural geologists, on the other, although among the latter there are a few important exceptions. Each side maintains its position confidently and tries to explain away conflicting evidence presented by the other. On the whole, however, the case presented by the paleontologists is the more convincing because their findings do not appear to be as liable to errors of observation and interpretation.

The evidence of the fossils has been countered in two ways. First, it is suggested that they are not naturally indigenous to the salt-marl strata but were introduced much later into plastic or porous beds. Although they are found in every sample studied, this explanation would appear to be possible if they were confined to the rock salt. However, they also occur in dolomite and in fine-grained oil shale, some of which consists of more than 50 per cent organic material. All geologists who have studied the area admit that the oil shale is an undoubted original constituent of the Saline series.

All who have examined the organic material agree that it includes fragments of angiosperm and gymnosperm plants and insects that must be of Cretaceous or younger age. If they are original constituents of the rocks—and it is almost inconceivable that they could have been

introduced into the oil shale—the age of the strata is established as post-Paleozoic. Gee's unconvincing secondary defense is the suggestion that these fragments represent an early flora of much more advanced type than anything previously found in beds of comparable age.

The proponents of an Eocene age for the salt marl counter the claim of the field geologist that no widespread thrust occurs above the Saline series by reminding them that thrusting is evident at some places and that the presence of thrust planes unreflected by any recognized local deformation has been established by geologists in several other parts of the world.

The printed symposium consists of nineteen papers, several of them quite inconsequential, and a section on discussion. Seven of the papers were read before the conference, eleven are by authors unable to attend but were briefly reviewed at that time, and three are subsequent contributions. Four papers are devoted principally to descriptions of fossil fragments obtained from the Saline series. Two papers consider theoretical aspects of Salt Range tectonics and support opposite conclusions. Among the more interesting and important geologic contributions are the following: "Microfossils and the Salt Range Thrust," by B. Sahni; "Some Tectonic Aspects of the Problem," by E. Lehner; "A Note on the Saline Series of North-western India," by M. S. Krishnan and N. K. N. Aiyengar; "Further Note on the Age of the Saline Series of the Punjab and of Kohat," by E. R. Gee; and "Further Notes on the Age of the Salt Range Saline Series," by E. S. Pinfold.

J. M. W.

An Introduction to Crystallography. By F. C. PHILLIPS. New York 3: Longmans, Green & Co., 1946 (July 2, 1947). Pp. ix+302; figs. 500. $6.50.

Part I, constituting three-fourths of this handy volume, is devoted to morphology; of the remainder, two chapters (x and xi) deal with space lattices and space groups and their symmetry, and a final chapter covers crystal habit. Physical (including optical), chemical, genetic, and X-ray crystallography are not treated. The nine chapters of Part I cover the study of crystals by the methods of single-circle reflection goniometry and the stereographic projection.

Chapter iv is a general study of the seven crystal systems in the order cubic, tetragonal, orthorhombic, monoclinic, triclinic, hexagonal, and trigonal; it deals only with the holosymmetric classes (also the pyrite class). The first third of the book could thus serve for an elementary course based on the stereographic projection (chap. ii), with brief treatments of goniometry and crystal drawing (axial cross). The remainder of Part I includes chapters on goniometry, the thirty-two classes (covered in the order of increasing symmetry), composite crystals, mathematical relationships, and crystal drawing. This sort of double coverage of crystal morphology in Part I is handled very successfully. It appears to be pedagogically sound and should result in imparting a good basic concept of the subject.

While it is true that "the main problems of [morphological] crystallography can be solved . . . . by the use of the simple single-circle goniometer," it does not follow that this is the best modern training technique. Thus Barker states that "as much work can be effected by the two-circle instrument in four hours, as in seven hours devoted to single-circle goniometry." Many feel that the two-circle machine, though more complicated to manufacture, is nevertheless much simpler to use. Moreover, the elegance of the associated gnomonic projection (very briefly treated in the present work) is of considerable educational value. This is indirectly suggested on pages 90, 219, and 276, in connection with the reading of face indices, crystal drawing, and the computation of reticular areas. The use of the Goldschmidt method in the new Dana's System is another important factor in favor of this projection.

Chapters x and xi serve as a very good introduction for beginning students to Volume I of the *International Tables for the Determination of Crystal Structures*. The Hermann-Mauguin notation is employed. Space groups are covered in the same order of crystal classes as is given in chapter iv, except that the tetragonal precedes the trigonal. In the final chapter the meaning of the Law of Bravais as generalized by Donnay and Harker is made clear.

The book has remarkably few errors; none of a typographical nature was noted. The definition of a crystal "as a homogeneous solid bounded by naturally-formed plane faces" is applicable only to that minority occurring as euhedrons. The "linear" projection (p. 31) should be called the "euthygraphic"; the former term was first applied to the Quenstedt projection. Hauy's *Traité* has five volumes, not four

(p. 34). The "stroke" in the symbol $2/m$ is called a *virgule* (p. 105). Re-entrant angles (p. 163) are typical of aggregates; a small proportion of these may be twins. The $c < a < b$ rule is not mentioned; thus the orientation of chrysoberyl (p. 170) does not fit Dana's System. The same holds true for topaz (p. 287), which in the new orientation to be adopted in Dana will not be prismatic (as stated on p. 286). Moreover, since it is said not to be holosymmetric (p. 175), the argument on pages 285–86 is vitiated on two counts.

The treatment in this book, greatly aided by numerous line sketches, is logical and lucid; the student should enjoy using it. The problems worked out in detail, together with the careful goniometric instructions, make it a real guide for the study of *crystals* (rather than models). It deserves to succeed, measured (as the author suggests) "by the number of its readers who finally lay it aside and 'throwing off the shackles of the text-book,' set out upon their own crystallographic investigations."

D. J. F.

American Oil Operations Abroad. By LEONARD M. FANNING. New York and London: McGraw-Hill Book Co., 1947. Pp. 270+vii. $5.00.

Treating his subject from the point of view of an oil company, Mr. Fanning has covered American petroleum operations in foreign countries in a most interesting manner. One must commend the remarkably profuse and well-chosen photographs with which the work is filled.

It has been the reviewer's contention that the public should be apprised of the risks in foreign oil fields. Mr. Fanning has presented effectively the undesirability of attempting such operations without capital running into many millions of dollars. His chapter dealing with political risks stresses the constant danger of expropriation and revision of terms. Economic risks are given almost equal space, with special attention paid to the time lag in dividends. He does not, however, point out the employmental, geological, and technological risks involved.

The chapter on the advantages, to both nations, covers social, educational, and economic factors. Although one may question some of the general premises of his economic discussion, the main points are clearly stated for the benefit of the American public. His picture of educational and social life in foreign oil fields is one unfamiliar to the average citizen. It might well receive more general consideration, with advantage both to individual and industry.

Curiously enough, Mr. Fanning has failed to distinguish between oil companies operating from an American and those operating from a European base, either in regard to their acceptance by, or in their treatment of, the nationals of the concession-granting countries. This reviewer thinks that he has missed an opportunity to point out that, on the whole, the American is better received and has given a better account of himself abroad than has the European.

In the appendixes one finds a vast amount of material; United States and foreign investments, production, price, trade, employment, and other statistics are available galore. They are not, however, broken down by countries, which renders them less usable than they otherwise would be. Furthermore, stress has been laid upon investment and financial considerations rather than upon physical quantities.

For the general reader or as collateral reading in general courses in college the book is admirably adapted. Both Mr. Fanning and the publishers are to be congratulated upon having presented a concise account of the search for oil in foreign lands by American nationals.

From the point of view of the research worker, however, the book has shortcomings. This reviewer is surprised to find a regrettable dearth of information relative to the geology or operating techniques. Nowhere is there a definite statement of the areas held in various countries or the underlying reserves.

H. W. STRALEY III

Rollin Thomas Chamberlin

Professor Emeritus of Geology and Editor Emeritus of the Journal of Geology

BORN OCTOBER 20, 1881
DIED MARCH 6, 1948

AS HAS been aptly said, an institution is commonly but the lengthened shadow of one man. So it was for a quarter of a century with the Journal of Geology. Rollin Thomas Chamberlin gave a considerable part of his life to the Journal and, by his service to that publication, rendered a large service to geologists and the science of geology. His association with the Journal exceeds in length even that of T. C. Chamberlin, its founder and first editor. Rollin Chamberlin was made a member of the editorial staff in 1912; he became managing editor in 1923 and editor in 1929. He remained in this capacity until his retirement on June 30, 1947. Readers and authors alike owe Rollin Chamberlin a great debt for this thirty-five years of painstaking and at times tedious labor. Many readers, perhaps, are unaware of the patient and unending effort contributed gratuitously by the editor in assisting authors in the preparation of their manuscripts for publication.

Rollin Chamberlin's service to the Journal did not consist solely in discharge of his editorial duties. He was also the author of many articles and reviews. His first contribution, "The Glacial Features of the St. Croix Dalles Region," appeared in volume 13 in 1905. His last paper, "The Moon's Lack of Folded Ranges," appeared in volume 53, 1945.

It is the hope of the present editorial staff that the Journal of Geology will continue as a publication worthy of the Chamberlin tradition.